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# Global Turbidity Studies. I. Volcanic Dust Effects— A Critical Survey

D. Deirmendjian

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**10. ABSTRACT**

A Critical evaluation of the role of the volcanic dust introduced by three major eruptions--Krakatoa (1883), Katmai (1912), and Agung (1963)--in increasing atmospheric particulate turbidity. Typical turbidity anomalies, expressed as absolute increments in optical thickness in the middle of the visual spectrum, are found to be 0.55 for Krakatoa, 0.35 for Katmai, and 0.25 for Agung. The last represents a fivefold increase of normal turbidity away from cities over a period of 2 to 3 years. No evidence of climatic effects directly related to the volcanic dust incursions is found. The possible contribution by the operation of 500 commercial supersonic transport vehicles to particulate turbidity is estimated to be small and climatologically not significant. An initial "black cloud" experiment, consisting of a simple reduction by 10 to 20 percent of the incoming short-wave radiation, is suggested for use with numerical models of the general circulation of the atmosphere to simulate volcanic dust effects.

**11. KEY WORDS**

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## PREFACE

Meteorological studies suggest that technologically feasible operations might trigger substantial changes in the climate over broad regions of the globe. Depending on their character, location, and scale, these changes might be both deleterious and irreversible. If a foreign power were to bring about such perturbations either overtly or covertly, either maliciously or heedlessly, the results might be seriously detrimental to the security and welfare of this country. So that the United States may react rationally and effectively to any such actions, it is essential that we have the capability to: (1) evaluate all consequences of a variety of possible actions that might modify the climate; (2) detect trends in the global circulation that presage changes in the climate, either natural or artificial; and (3) determine, if possible, means to counter potentially deleterious climatic changes. Our possession of this capability would make incautious experimentation unnecessary, and would tend to deter malicious manipulation. To this end, the Advanced Research Projects Agency initiated a study of the dynamics of climate to evaluate the effect on climate of environmental perturbations. The present Report is a critical evaluation of the particulate turbidity introduced by major volcanic eruptions and future supersonic transport systems; it is intended to assist in the formulation of reasonable models for use in numerical simulations. Related Rand studies by the present author include R-590, *Use of Scattering Techniques in Cloud Microphysics Research. I. The Aureole Method*, and *Electromagnetic Scattering on Spherical Polydispersions*, American Elsevier, 1969.

# VOLCANIC DUST EFFECTS

## SUMMARY

On the basis of available data this Report critically evaluates the role of the volcanic dust introduced by the three major eruptions of Krakatoa (1883), Katmai (1912), and Agung (1963) in increasing atmospheric particulate turbidity. Typical turbidity anomalies, expressed as absolute increments in optical thickness in the middle of the visual spectrum, are found to be 0.55 for Krakatoa, 0.35 for Katmai, and 0.25 for Agung. The last represents a five-fold increase of normal turbidity away from cities over a period of two to three years. The contribution by the operation of a nominal 500 SST commercial vehicles to particulate turbidity is estimated to be small and climatologically not significant.

No evidence of climatic effects directly related to the volcanic dust incursions is found. An initial "black cloud" experiment, consisting of a simple reduction by 10 to 20 percent of the incoming short-wave radiation, is suggested for use with numerical models of the general circulation of the atmosphere.

## CONTENTS

# 1. INTRODUCTION

In trying to evaluate the effects of global atmospheric turbidity on weather and climate, it would appear useful to review critically the available information on anomalous, long-lived changes of turbidity over large portions of the earth in order to estimate the amount and nature of the turbid components in each case. This information in turn may be correlated with the nature and magnitude of the climatic effects, if any, in an effort to understand the responsible mechanism so that eventually one may predict future effects with some confidence.

One type of large turbidity anomaly, unmistakably identified as such, is that produced by recorded extraordinary volcanic explosions capable of injecting massive quantities of so-called volcanic dust into the lower stratosphere and thereby of altering considerably its normal optical properties. It would then be a matter of recording the results of measurements--carefully obtained by proper techniques before, during, and after volcanic dust periods--of such optical parameters as the magnitude and the wavelength dependence of the scattering and absorption coefficients of a sample of turbid air, as well as the angular distribution of the intensity and degree of polarization of the scattered light. Such a collection of data, when analyzed in terms of a theory capable of explaining the scattering mechanism as a function of the shape, nature, and size of the scattering particles, would yield the desired information on the nature and amount of volcanic dust.

Unfortunately, as seen below, neither of these prerequisites were completely met during the most recent onset of volcanic dust. Consequently, whatever climatic variations might be attributable to the presence of such dust can only be explained rather qualitatively and with little confidence in terms of the anomalous turbidity produced by the volcanic particles.

Specific examples of this situation are three volcanic events that occurred within the last one hundred years and produced widespread turbidity anomalies, viz., Krakatoa (1883), Katmai (1912), and Agung (1963). The first was the subject of a very detailed record a·

analysis of all reported geophysical phenomena, published only five years after the main event [Symons, 1888].* Among these, the pertinent optical phenomena were recorded by means of qualitative and subjective visual observations, and their interpretation was based on an incomplete understanding of the scattering process. The second event, Katmai in Alaska, was not the subject of special study. However, some quantitative evidence of anomalous turbidity is available in the records of the Smithsonian Astrophysical Observatory, which was monitoring the solar constant on a routine basis at that time.

The most recent event, the 1963 eruption of Agung volcano in Bali, because of a combination of circumstances, did not receive the scientific attention it merited. For example, no national or international effort was made to organize *systematic* observations and measurements from strategic locations around the globe of the anomalies in atmospheric spectral transmission, and in sky brightness and polarization that followed this well recorded event for at least two years. If available, such data--once analyzed with the help of the more accurate theoretical models now available--would allow not only for a better assessment of the nature and extent of the volcanic dust layer, but also of its changes with residence time in the atmosphere. Instead, with a few exceptions, all we have are haphazard reports of visual observations of unusual optical effects by the occasional observer endowed with old-fashioned curiosity. Possible explanations for this lamentable loss of the fruits of a splendid and rare natural experiment are not difficult to suggest: A general decline of scientific interest in *natural* events capable of significantly altering our environment in the face of an increasing human potential of doing so by design or accident; and the related reluctance to devote much time to understanding our natural environment while engrossed in and fascinated by the new field of space exploration and experimentation.

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*Hereinafter referred to as "K.R." References are listed alphabetically by author and year of publication, pp. 55-58.

Nevertheless, it is worthwhile to re-examine the existing body of information in order to arrive at more reliable judgments about the nature of volcanic dust layers on the basis of present theoretical and experimental knowledge of light scattering phenomena. This we attempt to do here, with a view to eventually setting up a working model of the volcanic-dust-laden atmosphere for testing various conjectures about climatic effects.

## 2. AN EVALUATION OF THE KRAKATOA REPORT

By far the most comprehensive study, up to that time, of any single volcanic event was undertaken by the thirteen members of the Krakatoa Committee of the Royal Society, whose full report was published in 1888 under the editorship of the Committee's Chairman, G. T. Symons, F.R.S.* This document, composed with outstanding clarity and style, still makes instructive reading even in the light of up-to-date knowledge [Deirmendjian, 1957].

For the sake of completeness, let us recall briefly that the paroxysmal explosion of Krakatoa volcano, on the island of the same name, located in the Sunda Strait at about 6°05'S 105°30'E, occurred over the 26 and 27 August 1883; that several ships were within a radius of 100 miles from the site at the time of the major event [K.R., pp. 1-29] allowing eye-witness observation of the initial stages of the dust cloud; and that this is the explosion mainly credited with the injection of the large mass of particulates into the lower stratosphere, thus producing the optical phenomena later observed worldwide.

The volcanological data and other geophysical phenomena accompanying the main Krakatoa event, discussed in Parts I, II, III, and V of the Symons report [K.R., pp. 29-151, 465-475] is not touched upon here, for reasons of economy. We concentrate instead on the meteorological optical phenomena discussed in Part IV [K.R., pp. 151-426], which are our main concern.

### 2.1 Description of the Optical Phenomena Related to Krakatoa

The optical phenomena related to Krakatoa may be classified into three significant types: blue or green color of the sun or moon, Bishop's rings, and unusual twilights. Each is briefly described below, in that order (rather than the order followed in the Report); i.e., according to the increasing complexity of responsible mechanism. These phenomena did not necessarily appear simultaneously or seriatim

*The then President of the Royal Society, G. G. Stokes, was also a ranking member of the Committee.

at one spot, nor did they follow any particular order in time or space; but all three were undoubtedly connected with the Krakatoa dust.

2.1.1 Blue or Green Sun and Moon. The significant facts, as they emerge from the Krakatoa Report, may be summarized as follows:

a. The phenomenon was *visual* and *subjective* but seen by many reliable observers. Undoubtedly on such occasions the broad spectral continuum of the directly transmitted sunlight differed from that under "normal" conditions, giving the visual impression of a greenish or bluish hue; but data on the spectral transmission of the atmosphere were not available. The few solar spectra mentioned are of little help and difficult to interpret as they are not properly calibrated as to wavelength and absolute intensities. The hoped for discovery of some exotic volcanic *gas* responsible for the phenomenon did not materialize and it was agreed that the general "cutoff" observed toward the red portion of the spectrum must be attributed to solid particulate material.

b. The phenomenon was seen mostly in the tropical zone around the equator during the first few weeks immediately following the main eruption, and only rarely and with less reliability outside the tropics.

c. The blue coloration was observed mostly when the sun reached an elevation of at least 10° or so above the horizon and even near culmination; whereas, at low elevation near sunset or sunrise, the disc's color tended to be green, "yellowish-green," or "yellowish-white." When the sun set green, the rising moon was also greenish and so were bright stars and planets near the horizon.

The Krakatoa Report mentions other circumstances that might be significant: blue and green suns were observed together with unusually red twilight skies; a large sunspot was seen by naked eye on the green sun just before sunset; the fully eclipsed moon was observed to lack

the copper tint usually produced by earthshine. Also noted are other reports of similar phenomena: e.g., blue suns had been observed through Sahara dust in the air; blue and green suns had been reported in connection with volcanic events occurring in 1821, 1831, and other years.

The blue and green sun reports under Krakatoa dust conditions are significant because the phenomenon is indeed uncommon (cf. the expression "once in a blue moon"). It is not uniquely attributable to volcanic dust, since it has been observed also through other aerosol layers; conversely, not all volcanic dusts have produced blue and green sun effects. These effects may be interpreted in terms of simple attenuation (through scattering and absorption) by primary scattering without considering directional scattering features.

2.1.2 Bishop's Ring. This is one of the more significant phenomena associated with the Krakatoa dust, correctly called "a large corona" in the report. It is named after the Rev. Sereno E. Bishop, of Honolulu, who first described it in print, after his initial observation of the phenomenon on 5 September 1883, just ten days after the Krakatoa explosion. It was subsequently reported by several observers from various geographical locations for at least two-and-a-half years. From the descriptions carefully collected by the Committee [K.R., mainly pp. 232-263], the following significant features may be noted:

a. The Bishop's ring phenomenon appears to have been an aureole-corona complex within a circular region around the sun (or moon) of a visual radius 20° to 30°.

b. The name was applied more specifically to a system of corona rings, more or less colored so that a reddish outer ring of an average measured radius 22° to 23° surrounded a bluish white inner ring around the sun, whose border had an average radius of about 10.5°. The "inner radius" of Bishop's ring is necessarily imprecise, depending on the definition of the "inner border" mentioned in the Report. However, from the descriptions,

and as mentioned in the Report, the phenomenon was distinctly different from the usual 22° halo (in which of course the order of the colors is reversed) known to be the result of refraction through hexagonal ice crystals; rather, it conformed to the smaller diffraction coronas often observed around the sun and moon through optically thin clouds. (There was even a report of the appearance of a common halo with sundogs concurrently with a Bishop's ring.)

c. The phenomenon was definitely connected with matter in the higher atmosphere, since it appeared to be more brilliant when observed from high mountains in otherwise clear air, away from large cities, and it was independent of the presence or absence of cirrus clouds.

Additional interesting features are that the ring was seen around the moon as a "red haze" with a deep red outer border; that no change in the overall size was observed during the first 12 months of its appearance but was most prominent about 8 months after the eruption; and that there was an apparent eccentricity of the sun toward the horizon when Bishop's ring was observed near sunset.

A quotation from the *Comptes Rendus* by the eminent French physicist Cornu, whose observations should be highly reliable, describes the order of the colors in Bishop's ring to be the same as in the primary corona produced by thin clouds. Cornu also remarks that the phenomenon was most distinct when the scattering from lower atmospheric haze was "eliminated" (i.e., minimized) by the shadow of some cumulus cloud [K.R., p. 252].

To quote from the report, the Bishop's ring appears to have been a phenomenon "unique in the annals of optical meteorology . . . for size, brilliancy, universality, and protracted duration" [K.R., p. 247]. Unfortunately, such *optical meteorological* observations have gradually ceased to be considered important; so that in these times of almost exclusive reliance on instruments and quantitative analysis, we are apt to miss some significant intelligence on unusual meteorological conditions.

At any rate, the importance of the Bishop's ring phenomenon lies in providing the means to determine with reasonable confidence the size distribution and predominant size of the volcanic particles without recourse to higher order scattering theory.

2.1.3 Unusual Twilights and Crepuscular Phenomena. The descriptions of this aspect of the post-Krakatoa skylight effects, although among the most widespread and noticed, are perforce neither as explicit nor as uniform as those of the blue sun and Bishop's ring. This is due to the close dependence of twilight features on the optical characteristics of the local and trans-horizon troposphere, and their variation not only along the vertical but along the horizontal direction as well; and to the highly subjective nature of the individual impressions of twilight colors, relative brightness, and their changes with time.

The salient aspects of the post-Krakatoa twilights may be listed as [K.R., pp. 152-178]: their unusually long duration and vivid coloration, both said to be markedly different from "normal" twilights; the presence of an unusual "diffuse illumination" of the whole sky so that the earth shadow could not be distinguished; and the existence of a secondary twilight glow clearly separated from and at a higher elevation than the primary or "ordinary" glow usually seen above the sunset horizon on clear days. This secondary glow could be distinguished by a characteristic delicate purple hue rather difficult to describe verbally.*

There were reports of other phenomena observed in the cloudless daytime sky, such as unusual skylight colors away from the sun, a considerable diminution in the polarization of skylight, and the appearance of several new neutral (zero-polarization) points outside the sun's vertical plane.

*Anyone who has not been fortunate enough to have seen the rather similar post-Agung twilights (see Sec. 4.1), as the present author has, will be quite unable to visualize the delicate coloration that these observers were trying to describe.

All these phenomena (except those of polarization) were observed by numerous people at various widely separated locations and for long periods of time after the main eruption. The great number of visual descriptions, collected so carefully by the Krakatoa Committee, must be attributed to the general scientific curiosity characteristic of the times and perhaps to the absence of movies, radio, television, etc. A critical observation of Nature had not yet become the exclusive domain of professional scientists. The veracity of the descriptions must not be underrated, especially when made by sailing-ship officers who were keen observers of atmospheric optical phenomena (before the advent of the professional meteorologist) for their weather-forecasting value.

The bulk of the evidence presented in the Krakatoa Report supports the assumption that most of these phenomena were definitely connected with the volcanic dust. In fact, the systematic logging of the progressive appearance of the optical phenomena over different parts of the globe provided the first evidence of a high-level equatorial ring of strong winds that distributed the volcanic matter at such a fast rate. The Report estimates that about 26 days after the paroxysmal eruption of 26 August 1883, the dust had made two complete circuits around the globe traveling from East to West. The rate of travel of the dust cloud was estimated at between 63 and 72 knots! The discovery of these sometimes called "Krakatoa winds" near the 30-km level is one of the main contributions of the Report. Their systematic study has been initiated only within the past decade or two.

## 2.2 Interpretation of Post-Krakatoa Optical Effects

From the mainly qualitative nature of the observations, it follows that an interpretation of the optical phenomena in terms of the Krakatoa dust cannot be very specific and quantitative. We here outline some conclusions on the basis of the phenomena described above, considered both separately and collectively. These conclusions do not necessarily reflect those reached by the authors of the Krakatoa Report.

First, two general remarks independent of a detailed analysis of the phenomena: (1) The dust particles cannot have been strictly monodisperse and uniform in composition and shape, otherwise the phenomena should have been much more spectacular than reported [cf. Deirmendjian, 1969a;* this follows also from their origin and mode of formation]. (2) The chemical composition of the particles, after a considerable residence in the lower stratosphere, may have changed from that of the original volcanic material, due to photochemical or other local reactions.

2.2.1 Blue Sun. In attempting to explain the blue and green suns, the most straightforward hypothesis is that the spectral extinction provided by the Krakatoa dust is responsible for the changes in spectral composition of the directly transmitted parallel sunlight needed to produce the visual impressions described in 2.1.1(a). By extinction, of course, we mean the depletion produced by both scattering and true absorption.

Under this assumption, by applying the well-known Bouguer-Langley principle to the "observed" overall spectral transmission, one may derive the extinction optical thickness, $\tau(\lambda)$, of the volcanic dust-laden atmosphere as a function of wavelength in the visual range. From this, in turn, one may deduce the volcanic dust component, $\tau_D(\lambda)$, by subtracting the parts contributed by the molecular or Rayleigh atmosphere, $\tau_R(\lambda)$, and the "normal" aerosol component, $\tau_M(\lambda)$, for otherwise clear days. It is then a question of constructing a volcanic-dust model endowed with optical constants (i.e., composition), size range, and size distribution such that it will yield a theoretical optical thickness closely approximating $\tau_D(\lambda)$ in magnitude and spectral variation.

The trouble is that visual color impressions, such as those reported in the Krakatoa Report, are notoriously unreliable indicators of the true spectral radiance, especially when, in the case of the sun, the impressions were recorded against the existing skylight background. Similarly, in the case of the moon and Venus, we have to consider subjective impressions recorded against the memory of the "normal" appearance of these bodies.

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*Hereinafter referred to as ESSP.

Nevertheless, it is clear that the Krakatoa dust must have depleted the red end of the solar spectrum by an amount greater than the normal atmospheric aerosols, as was essentially surmised in the Krakatoa Report. The question is by how much and how, i.e., whether by increased scattering or absorption. Even assuming polydispersions of homogeneous spherical particles, the question of choice is not an easy one--as may be seen from theoretical model investigations in this area [ESSP]. Because of the short spectral range involved (visible) and the subjectivity mentioned above, there may be several models, with different compositions and size distributions, yielding the same type of spectral attenuation. However, taking into account the other optical phenomena mentioned, we may be able to make an educated conjecture on the general characteristics of the Krakatoa dust--thus narrowing down the range of choices, which may be further limited by observations and other information from more recent volcanic-dust phenomena. It may then be possible to set up a synthetic model involving particulates of some hypothetical composition (or optical properties), not necessarily corresponding to any known substance but in fair agreement with observations otherwise.

At any rate, such analysis must be limited to a consideration of idealized, homogeneous, spherical particles--the only type whose scattering and absorption properties are known with any degree of completeness, either for individual particles [van de Hulst, 1957], or in polydisperse aggregates [ESSP]. Consideration of irregular and inhomogeneous particles is barred by a lack of complete scattering theories for such particles. This situation is not as bad as it might appear, however, since it is known that in the case of polydispersions of relatively small particles, and for such properties as forward scattering and extinction cross section, the shape factor is not important provided the particle's volume and optical constants are specified [cf. ESSP, p. 108; Holland and Gagne, 1970].

For the moment, let us require that the optical thickness of the volcanic dust be such that, when combined with that of the normal atmosphere, it should yield an overall value that is practically independent of wavelength. This condition will indeed result in a blue-

green visual impression of the sun when observed through a sufficient amount of volcanic dust, as described in 2.1.1(a) and (b), since the extraterrestrial spectral radiance of the sun has an absolute maximum in this region of the spectrum--i.e., around $\lambda 0.45\mu m$ [cf. Allen, 1958]--and the dusty atmosphere will merely act as a neutral filter.

Furthermore, the value of $\tau$ in the middle of the visible range, say at $\lambda 0.5\mu m$, should be several times the normal value, to enable the observation of a sunspot by the naked eye under the conditions mentioned in 2.1.1. These conditions must have been similar to those for the "filtered sun" effect mentioned elsewhere [ESSP, p. 114; Deirmendjian, 1969b], which are equiavlent to a slant-path transmission of about $e^{-16}$. Assuming that the sun was about 3° above the horizon when the sunspot was observed, and from the usual considerations about air mass being given by sec $\zeta_o$ ($\zeta_o$ = zenith distance of the sun), we have

$$16 \simeq \tau \sec 87^\circ$$

whence

$$\tau(\lambda 0.5\mu m) \simeq \frac{16}{19} = 0.84 \quad .$$

This is about four times a typical value of the normal optical thickness likely to be found at $\lambda 0.5\mu m$ in clear air away from urban pollution [Deirmendjian, 1970]. By the same token, then, the volcanic-dust contribution to the total optical thickness at this wavelength would be

$$\tau_D(\lambda 0.5\mu m) = \frac{3}{4}\, 0.84 = 0.63 \quad , \qquad (1)$$

i.e., about three times the total atmospheric value for a clear day. The value thus obtained in Eq. (1) is not really excessive if we consider the circumstance mentioned in 2.1.1(b), i.e., the implied high *initial* density of the dust within e volcano's latitude zone related with sightings of the blue sun. Hence, the value suggested by Eq. (1) should not be considered typical for the long-lived dust responsible for the more widespread phenomena.

Circumstance 2.1.1(c)--the "greening" of the blue sun when very near the horizon--is not difficult to explain qualitatively in terms of the earth's sphericity and the high level of the volcanic-dust layer (the Chapman effect): it has to do with the fact that the extinction contribution by the lower and denser atmospheric layers increases relative to that of the higher layers as the sun approaches the horizon, because of the geometry of the situation. Since the troposphere was relatively free of the dust, and its normal effect is that of reddening a setting sun, the resultant effect on the blue sun would be a greening tendency. This hypothesis, which is similar to the one proposed by the Krakatoa Report, cannot be tested quantitatively without actual measurements of spectral transmission of the setting blue and green suns.

2.2.2 Bishop's ring. Accurate data on the size and order of colors in the various rings, as well as the brightness gradient in this phenomenon would have been most useful in estimating the size and size distribution of the dust particles. Unfortunately, the only quantitative type of information available is the radius of such subjective features as the inner or outer "border," "middle," or "brightest part" of the red ring, etc., which are of little help without absolute or even relative narrow-band photometry.

If monodisperse or quasi-monodisperse particles are assumed, the Report's original estimate [K.R., p. 257] of a radius of about 0.8μm, on the basis of *first-order diffraction theory for opaque discs*, may not have been too bad, all things considered. To account for spherical particles (rather than thin discs), this estimate would have to be revised downward by about 25 percent [van de Hulst, 1957; ESSP, pp. 50-53], which yields 0.6μm for the radius of such idealized Krakatoa particles.

However, one cannot reasonably assume that the volcanic particles were strictly monodisperse, for then the associated phenomena would have been much more spectacular in color and brightness. On the other hand, the size distribution cannot have been very wide, for then there could be no Bishop's or other corona-type rings but only a "white"

and featureless aureole [ESSP, p. 94ff]. Nor can the modal size and type of distribution law have been the same as those in common, corona-producing tropospheric clouds [cf. ESSP, p. 107], which produce coronas too small in radius (1° for the inner to 12° for the outer border) and which involve particles too large (2μm to 4μm modal radius) for the necessary long stratospheric residence. We may thus assume that the Krakatoa dust conformed to an unusual type of relatively narrow size-distribution law around a modal radius near 0.6μm, which, when combined with the normal tropospheric aerosol distribution, produced the Bishop's ring phenomenon as reported.

2.2.3 Unusual twilight phenomena. Because these phenomena are greatly affected by secondary mechanisms (e.g., second- and higher-order scattering and indirect illumination), and they depend on a number of parameters (e.g., local extinction coefficient as well as directional scattering) and their variation both in the vertical *and* in the horizontal direction, the post-Krakatoa twilights cannot, in general, be interpreted in a straightforward manner.* As pointed out by Rozenberg [1963], even if accurate quantitative data were available, the information deduced from their analysis could hardly be unique or unambiguous.

The main significance of these unusual twilights lies in their indication of the great height of the responsible dust layer, inferred from their long duration and the presence of secondary glows; in their unusually high brightness, indicating that even if the dust were composed of light absorbing material, it nevertheless was characterized by a high albedo of single scattering; and in their visibility from locations distributed over a wide latitude zone (where the other optical phenomena were rarely, if ever, observed), indicating the wide spread of the volcanic dust over higher latitudes within both hemispheres.

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*Similar remarks apply to the considerable anomalies in skylight polarization phenomena observed in the presence of volcanic dust [Sekera, 1957]. This subject is not discussed here for lack of space.

## 2.3 Other Meteorological Effects Related to the Krakatoa Event

For those interested in atmospheric effects such as unusual deviations of surface temperature, pressure, winds, cloudiness, precipitation, etc. from the norm (i.e., weather or climatic effects), there is very little useful information to be found in the Krakatoa Report. A section on the general analysis of unusual meteorological phenomena [K.R., pp. 426-463] is confined to proving that the *optical* phenomena are directly related to Krakatoa's dust and to refuting contrary arguments. Only as a by-product, and by following the dates of first appearance of the optical phenomena, is a general description of the equatorial and sub-tropical stratospheric circulation deduced for the first time.

Some additional interesting conclusions and conjectures, arrived at by the Krakatoa Committee, are worth mentioning. One is the initial estimate of the mass of material contained in the airborne dust, equivalent to some 4 cubic kilometers of "solid matter expelled from the volcano" out of a total of 18 $km^3$ for all the ejecta [K.R., p. 440] and a suggestion that this might be revised upward by a factor of 4 or 5 under certain assumptions [K.R., p. 448]. Another, even more interesting point, is the suggestion that the main portion of the stratospheric dust might have been composed of "condensed gaseous products of the eruption (other than water) such as sulphurous acid or hydrochloric acid" [K.R., p. 445]. In the light of contemporary ideas about the origin of stratospheric aerosols [cf. Junge *et al.*, 1961] and their *in situ* photochemical generation, the above suggestion may not be too far-fetched.

# 3. THE KATMAI EVENT OF 1912

## 3.1 General Description

On 6 June 1912, a major convulsion apparently culminated in the blowing-off of the top of Mt. Katmai, on the Alaska Peninsula near the Shelikoff Strait, at about 58°16'N 155°W. Although perhaps comparable in intensity to Krakatoa, the Katmai event (unlike its predecessor) was not the subject of a systematic collection and analysis of observations of the directly related geophysical phenomena. This may be partly attributed to Katmai's remote location and the absence of subsequent spectacular skylight phenomena over populous cities.

As with Krakatoa, however, we have some description of the initial stages as seen by ships' captains who were in the vicinity of Katmai at the time. An example is Capt. K. W. Perry's report of the U.S. Revenue Cutter *Manning* [Anonymous, 1912] while bunkering at Kodiak, about 100 miles from the volcano, on the day of the explosion, in which he describes the copious ash-fall on the town (and subsequent damage and hardships to the population) and the reduction of visibility to practically zero for two days. When some visibility returned after 40 hours, "the skies (at 2:30 pm) assumed a reddish color."

A more comprehensive description of the main event was given by George C. Martin, a geologist with the U.S Geological Survey, who, though not an eye-witness himself, visited the site just four weeks after the eruption [Martin, 1913] and collected some eye-witness descriptions. We learn that the mail steamer *Dora* was actually steaming up the Shelikoff Strait on the evening of 6 June, and Captain C. B. McMullen was unable to make Kodiak harbor when visibility dropped to zero two hours before sunset ("not even the water over the ship's side could be seen") forcing him to run out to the open sea for 12 hours.

Several "natives" reported having seen Katmai with its upper half missing immediately after the explosion. (It remained for much later *ad hoc* expeditions to corroborate this to the satisfaction of non-natives.)

The *Bertha* reported seeing a red sun in "clear skies" about noon on 7 June when steaming about 375 miles NE of the volcano,

In comparing the Krakatoa and Katmai volcanic events, Martin [1913] conjectures that Krakatoa's must have been by far the greater of the two because its explosions were heard from as far as 3000 miles whereas Katmai's were not recorded farther than 750 miles. (This of course is not a valid criterion, due to differences in atmospheric structure such as tropopause height between equatorial and subarctic regions.) However, he also states that the two events were about equal in the quantity of material ejected.

More information about the nature and magnitude of the Katmai event was put together only after a number of later, semi-scientific expeditions were sponsored by the National Geographic Society, as described in the Society's magazine [cf. Griggs 1917, 1913, 1921]. These articles, prepared by the leader of the expeditions, the botanist Robert F. Griggs, abound with awe inspiring descriptions of the unexpected and fantastic wonders of nature produced by the volcanism but offer little help in evaluating, for example, the amount and nature of the material introduced into the atmosphere. However, from estimates of the shape and height of Mt. Katmai before the eruption and later surveys of the actual crater and caldera, Griggs arrives at an estimate of the total volume of ejecta of about 8 cubic kilometers (as against 18 for Krakatoa).*

On this basis, if we assume that the two volcanoes introduced the same type of dust into the atmosphere, and roughly by a similar mechanism, other things remaining equal, we must conclude that Katmai's contribution to long-lived atmospheric particulates must have been just under one half of Krakatoa's. This circumstance, together with

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*Note, however, that in a recent article, Ernest Gruening [1963] mentions conclusions reached by geologist G. H. Curtiss to the effect that the volcano actually responsible for the atmospheric--and other--volcanic ejecta was the newly formed Novarupta volcano located six miles west of Katmai, and that Katmai's top, rather than being blown off, collapsed within itself after the underlying magma was siphoned off by Novarupta. Also, according to this source, the total volume of ejecta was of the order of 29 cubic kilometers, and some material was initially injected as high as 40 km into the stratosphere. A similar uncertainty exists about Krakatoa's ejecta, with a possibility that the accepted figure may be an underestimate by a factor of four or five [K.R., p. 448].

the high-latitude location of Katmai within the prevailing westerlies, must explain the limited spatial and temporal extent of its atmospheric optical effects. Interestingly, the effects of the Katmai dust on atmospheric turbidity were first reported from the State of Virginia, 3700 miles away, then from Bassour, Algeria, at 6000 miles, and finally from Mt. Wilson, California, only 2500 miles from Katmai [Abbot, 1913].

## 3.2 The Katmai Turbidity

The Katmai turbidity effects may be evaluated thanks to the records of the then-operating Smithsonian Astrophysical Observatory, as published in its decadic Annals. The main objective of the Observatory, particularly of its director, C. G. Abbot, was the precise measurement of the solar constant and its possible temporal variations, hoping to eventually establish their relation with climatic variations. What is also known, perhaps not so widely, is that this attempt failed but, as a fortuitous by-product of the main objective, the records provided a unique and valuable source of material about atmospheric turbidity and its evolution. When the directorship of the Observatory changed hands about fifteen years ago, this program was discontinued despite its potential value [Deirmendjian, 1955a] in the above-mentioned sense. In effect, a record of the atmospheric spectral transmissivity, obtained by the Smithsonian's so-called "long method," and continuous observations from permanent sites over the last sixty years, would have been a most reliable index, not only of singular changes in turbidity due to major natural events and point sources, such as volcanoes, but of possible gradual changes that might be correlated with human activity.

Returning to the original program of the Smithsonian Astrophysical Observatory, since the solar energy flux could be measured only after its passage through the atmosphere, some means had to be devised to eliminate the effects of its variable transmissivity. This was accomplished mainly on the basis of two circumstances: (a) the then newly invented Langley spectrobolometer, capable of narrow-band flux measurements from the near-ultraviolet to the near-infrared

regions of the solar spectrum; (b) Langley's adaptation of the classical Bouguer-Lambert law to quasi-monochromatic solar radiation in the earth's atmosphere so that its transmission in the vertical direction and a relative extraterrestrial flux could be evaluated. To do this properly required as clear an atmosphere as possible and the assumption of rather stable optical properties during the needed series of observations as a function of the changing zenith distance of the sun. The absolute value of the solar constant was then deduced with the help of various corrections and an independent measurement of the total energy received with another instrument. The accuracy of this procedure will not be discussed here.

A close examination of the Smithsonian records for Mt. Wilson for the period following Katmai's eruption [Abbot *et al.*, 1922] clearly shows a general increase in turbidity that can very well be related to the volcanic dust. However, an identification of the component attributable to such dust, as distinct from other components of turbidity, is not easy since no effort in this direction was made at the time. The usual procedure [Deirmendjian, 1955b] is to determine the total optical thickness from the normal transmissivities listed in the tables, then subtract the permanent Rayleigh component and, in certain spectral regions, the estimated absorption due to ozone. The remainder, or *residual optical thickness*, may be attributed to turbid components such as natural and artificial aerosols in the form of particles larger than molecules and atoms. In particular, using the notation already introduced in Sec. 2.2.1, we may set

$$\tau(\lambda) = - \log_e q(\lambda) \qquad (2)$$

where $q(\lambda)$ is the recorded narrow-band transmissivity obtained by the Smithsonian's "long method;" whence we get

$$\tau_M(\lambda) + \tau_D(\lambda) = \tau(\lambda) - \tau_R(\lambda) - \tau_{O_3}(\lambda) \qquad (3)$$

where $\tau_R(\lambda)$ is the standard atmosphere Rayleigh component for the height of the particular station [Deirmendjian, 1955c]; and $\tau_{O_3}(\lambda)$

is the extinction produced by ozone absorption (e.g., in the Chappuis bands), depending on the amount above the station at the time of observation. Since the amount of ozone depends on location and season of the year as well as local weather situation, unless it is independently determined for the particular station, it may be evaluted only approximately in terms of latitude and time of year.

Evidently, the residual optical thickness, given by the left-hand side of Eq. (3) as the sum of two terms, can be a highly variable quantity, as it depends on many factors at any given locality. Unless very long records exist at any single observatory, which have been properly analyzed and classified, it is difficult to arrive at some value for normal background turbidity, $\tau_M(\lambda)$, that may be used to evaluate the anomalous component to $\tau_D(\lambda)$ due to volcanic dust or other unusual sources. With this proviso in mind, we try to estimate the Katmai dust component by using a reasonable value for $\tau_M(\lambda)$ based on clear-day data obtained before, and some time after, the volcanic dust presumably was over the Southern California coast.

On Fig. 1, the straight line marked $\tau_{1P}(1737)$ indicates the Rayleigh standard atmosphere optical thickness corresponding to the Mt. Wilson Smithsonian station elevation of 1737 meters above sea level, as a function of wavelength, excluding corrections for molecular anisotropy [Deirmendjian, 1955c]. The curves marked (a) and (b) are based on Mt. Wilson data, obtained by the general procedure described by Eqs. (2) and (3). In particular, the full line marked (a) represents the residual optical thickness ($\tau - \tau_R$) obtained by combining the recorded June 1912 (23 pre-Katmai-dust days) average transmission with that given for October 1913 (25 post-Katmai-dust days) in the Annals. Both these months must have been characterized by an exceptional number of very clear days, judging from the high transmission values recorded when compared to other individual months or days. We have chosen their average as a norm for clearest conditions at Mt. Wilson in those years, somewhat arbitrarily, in order to smooth out random variations at particular wavelengths without bias. (We also have performed no interpolations or other manipulation of the original data, but used the seven standard wavelengths 0.35, 0.40,

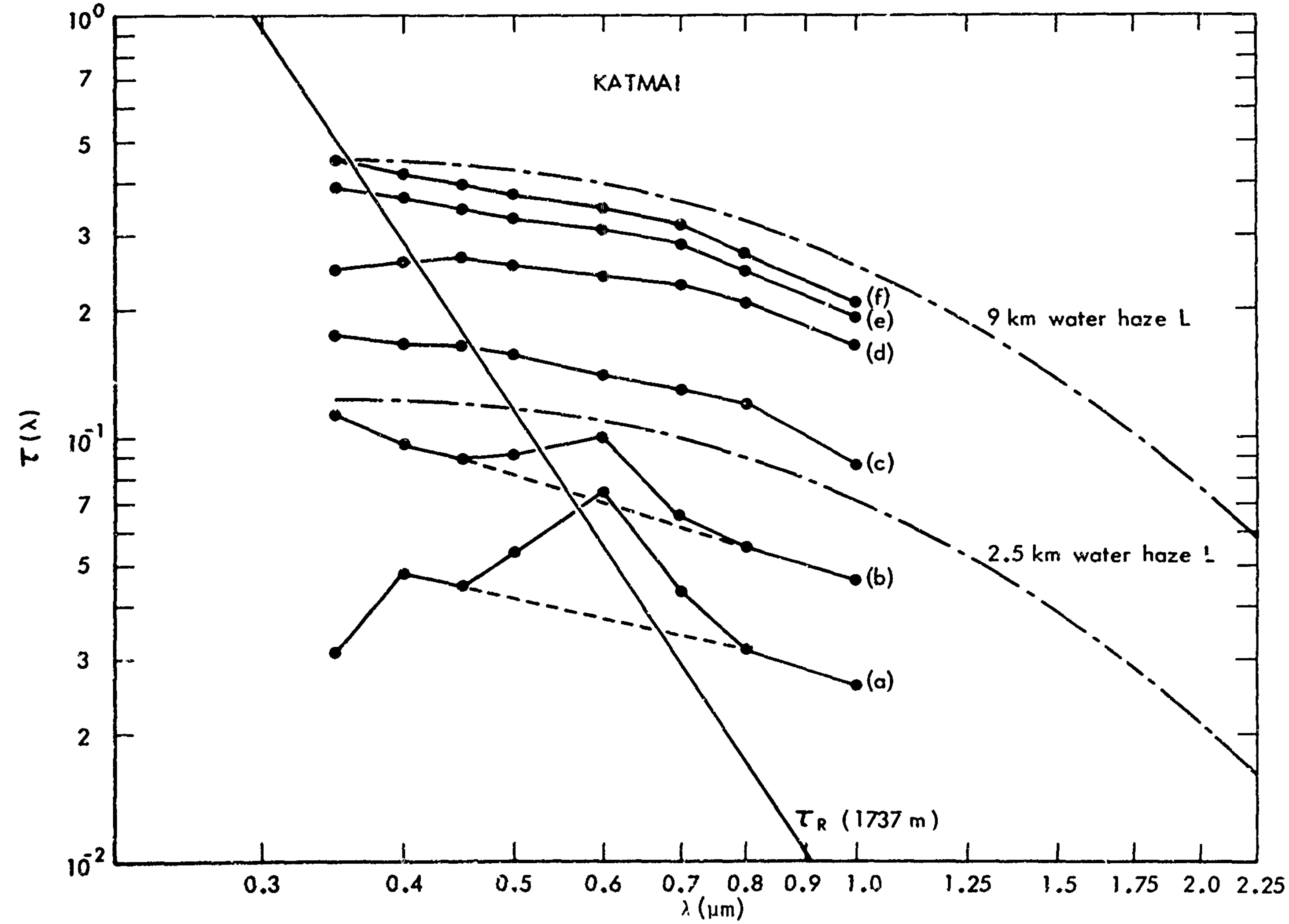


Fig. 1 -- The turbidity anomaly produced by the Katmai (1912) eruption, evaluated from contemporary data of the Smithsonian Astrophysical Observatory obtained at Mt. Wilson, California. (See text for explanation of curves.)

0.45, 0.50, 0.60, 0.70, 0.80, and 1.0μm given in the Smithsonian tables; and have joined the data points with straight line segments on the logarithmic chart of Fig. 1.)

Curve (b) corresponds to similar data but for 3 August 1913, a single day with relatively low turbidity over Mt. Wilson. The dashed straight-line portions joining the data points at λ0.45 and 0.80μm on curves (a) and (b) roughly indicate the effect of the ozone correction in the Chappuis continuum, which has a maximum absorption near λ0.6μm. Disregarding the portions for $\lambda < 0.35\mu m$ where the Smithsonian data could have been unreliable [Dunkelman and Scolnik, 1959], the single clear-day data represented by curve (b) and the 48-day average given in curve (a) are remarkably similar in shape and slope except for a factor that is almost independent of wavelength. This suggests that the nature and size distribution of "normal," clear-day aerosols over Mt. Wilson was constant and only the total concentration varied from day to day during the periods in question.

We could therefore consider conditions yielding a residual optical thickness represented by curves (a) and (b) in Fig. 1 as typifying the norm for very low turbidity over Mt. Wilson, and use them to evaluate the deviations during abnormally turbid conditions. In particular, such curves could be subtracted from corresponding ones for unusually turbid days and the difference might then be attributed to the anomalous turbidity, $\tau_D(\lambda)$, introduced by the responsible source on such occasions. An even simpler and more reliable procedure would be to subtract the total optical thickness (at each wavelength) obtained by means of Eq. (2) and corresponding to curves (a) and (b), from that derived from turbid-day data. Then, such corrections as for the ozone absorption [indicated by the area between the full and dash portions of curves (a) and (b)], and for possible instrumental and other errors, would automatically be taken into account, since all sets of data were presumably obtained with the same apparatus at the same location and reduced by the Smithsonian people in a standard manner. In the process, the almost invariant Rayleigh component will have been removed without any assumptions whatsoever about its theoretical magnitude.

This latter procedure was followed in obtaining the rest of the full curves in Fig. 1, labeled (c) to (f), from data on days when the Katmai dust had presumably arrived over the site. From these it is quite clear that the above method of reduction does indeed yield a set of remarkably smooth curves, all things considered, which seem to belong to the same family. For economy we do not show the minuends and subtrahends we used in obtaining the curves, but these were given by the optical thicknesses derived from the following data found in the original Smithsonian records [Abbot *et al.*, 1922]:

<u>Curve (c):</u> That derived from the average (arithmetic mean) transmission of 27 days in August 1912 *minus* that from 23 days in June 1912

<u>Curve (d):</u> That from the mean for 17 and 19 August 1912 *minus* that from the original data for curve (a)

<u>Curve (e):</u> That from data for 2 August 1912 *minus* that from the average of 19 days in August 1913

<u>Curve (f):</u> Also from data for 2 August 1912 *minus* the values from the original data for curve (a)

Although Abbot [1913] mentions that the first indications of Katmai dust were evident over Mt. Wilson as early as 21 June 1912, an examination of the monthly mean transmissions for that year, given in the Annals, reveals that the local turbidity was actually at a minimum in June, rose sharply during July and August, and probably reached a maximum during September or October (for which two months no average values are given). Our choice of the August 1912 days as representative of Katmai dust conditions, listed above and reduced as displayed in Fig. 1, was governed by this circumstance as well as by the grade or quality of raw data, recorded as "excellent" or "very good" in the Annals. These qualifiers were used to indicate that the corresponding Bouguer-Langley logarithmic plots, as a function of varying solar zenith distance, fall on straight lines, implying stable turbidity conditions and hence reliable values of atmospheric transmissivity.

Thus August 2, represented in curve (e) or (f), shows the highest Katmai dust content when compared to the lowest average turbidities, as in curve (a), found in June 1912 and October 1913. By the same criterion, the mean of two other days in August 1912, represented in curve (d), shows a Katmai dust turbidity equivalent to about two thirds of this maximum; and curve (c) obtained as above for the mean for all 27 days of August on which observations were made shows less than one half the maximum Katmai effect of August 2, but still almost four times the turbidity for very clear days as in curve (a). This gradual decrease as more days are used in averaging suggests that the Katmai dust layer arrived over Southern California in waves, as is very likely from present knowledge of upper tropospheric flow patterns in this region. These inferences, of course, must be considered with due caution in view of the known variability of local turbidity, which is produced by various meteorological conditions other than those connected with volcanic dust and is apt to result in turbidities as high or higher than shown in curve (f). We have no way, at this distance in time, of arriving at a low or moderate "norm" for the local turbidity, but the procedure outlined above should provide as reliable an evaluation as possible on the basis of the available data.

The family of four curves--(c), (d), (e), (f) in Fig. 1--derived as above from August 1912 data and interpreted as anomalous turbidity attributable to Katmai dust, can be bracketed very nicely by curves based on a single theoretical model that we had devised some four years ago (prior to the present analysis) to represent *light tropospheric turbidity* rather than any volcanic effects. This is demonstrated in Fig. 1 by the two continuous dash-dot curves, marked "2.5 km" and "9 km water Haze L," respectively, representing a corresponding thickness of homogeneous aerosol layer as given in the published data for this model [ESSP, pp. 77-79 and Tables T.16 to T.19]. Alternately, these two curves may be taken to represent the optical thickness produced by totals of $2.5 \cdot 10^7$ and $9 \cdot 10^7$ particles per $cm^2$ column, respectively, with an arbitrary variation of number density in the vertical direction, but a fixed size-distribution law, $n(r)$, called "Haze L," given by [ESSP, pp. 77-79 and Fig. 3 below]

$$n(r) \propto r^2 \exp\{-15.12\sqrt{r}\} \quad . \tag{4}$$

This model has a maximum concentration at a radius, r, of 0.07μm and a mixing ratio $V_p$ of $1.167 \cdot 10^{-11}$ [ESSP, p. 78]; so that, if the particles were to be precipitated and compacted into a bulk substance, they would occupy a volume of $2.92 \cdot 10^{-6}$ and $1.05 \cdot 10^{-5}$ cm$^3$ per cm$^2$ column above the surface, respectively, for the lower and upper bracketing curves. From the density of the particulate substance one could then determine their total mass.

Taking the upper estimate as a criterion, if we assume such a dust veil spread above the entire earth's surface or an area of some $5.1 \cdot 10^{18}$ cm$^2$, the total volume of the dust alone would be $5.36 \cdot 10^{13}$ cm$^3$ or $5.36 \cdot 10^{-2}$ km$^3$. This may be compared with Griggs' [1918] estimate of 8 km$^3$ for the volume of *all* the ejecta from the Katmai eruption. More realistically, if we assume the Katmai dust to have spread only over the spherical cap above 30°N latitude, the area covered would be only $\frac{1}{4}$ of the total, or $1.275 \cdot 10^{18}$ cm$^2$, which, for the same model, would require a dust amount equivalent to $1.34 \cdot 10^{-2}$ km$^3$ of solid material. This is a more reasonable assumption for the spread of Katmai's dust, considering the high-latitude location of the source and various other criteria; and it represents only $\frac{1}{600}$ of Griggs' above-mentioned estimate for the total ejecta.

To test the validity of this deduction, let us assume that the same fraction of Krakatoa's total ejecta, estimated at 18 km$^3$, was initially spread over the intertropical zone between 23°.5 North and South, representing $\frac{4}{10}$ of the earth's total surface. We get

$$\frac{18 \cdot 10^{15}}{600} \frac{1}{0.4 \cdot 5.1 \cdot 10^{18}} \text{ cm}^3$$

or about $1.47 \cdot 10^{-5}$ cm$^3$ of particles per cm$^2$ column. We then observe that on the basis of the same model as for the Katmai dust, this is equivalent to a 12.6 km thickness of homogeneous "haze L" layer, which, in the case of water substance, would result in a dust optical thickness, $\tau_D = 0.60$ at λ0.45μm. Surprisingly, this agrees rather well

with the estimate of $\tau_D = 0.63$ at $\lambda 0.50 \mu m$, given in Sec. 2.2.1, Eq. (1), and arrived at by an entirely different reasoning for Krakatoa's initial turbidity.

We note in passing that if we were to believe the Krakatoa Committee's own estimate of 4 $km^3$ of material injected into the atmosphere as dust, proceeding as above, we arrive at the equivalent of 1680 km of haze L, yielding a dust veil with the fantastic optical thickness $\tau_D \approx 80$ over the entire intertropical zone.

Finally, if one averages the $\tau_D$ values at $\lambda 0.45 \mu m$ for the Katmai turbidity, shown by the four curves (c) to (f) on Fig. 1, one gets 0.30, i.e., just under $\frac{1}{2}$ of the Krakatoa value mentioned above, in good agreement with the ratio estimated in Sec. 3.1 on the basis of the total amount of ejecta from the two volcanoes.

Note that our above estimates for the volcanic dust component, $\tau_D$, cannot change radically if the dust composition is changed, say, from water to some other nonabsorbing dielectric substance of higher density, *provided* the size distribution and total number of particles remain the same. This has been demonstrated by the theoretical models investigated by the author [ESSP] and others. It is thus interesting that the total mass of either the Katmai or the Krakatoa atmospheric dust (derived as above, assuming the density of water) could hardly have exceeded $10^{-8}$ of that of the entire atmosphere ($5.14 \cdot 10^{21}$ g) or the large portions thereof that were affected; whereas the corresponding total optical thickness was increased by more than a factor of two *with respect to cloudless and very clear conditions* away from urban pollution centers. The last proviso should be kept in mind when discussing possible climatological effects of volcanic turbidity.

# 4. THE AGUNG EVENT OF 1963

## 4.1 General Description

As mentioned in Sec. 1, the most recent volcanic event, the eruption of Agung--which is in the same class as those of Krakatoa and Katmai in the magnitude and extent of atmospheric effects--hardly received the attention it merited. Whatever useful data we have on related turbidity effects, for example, were randomly and accidentally obtained, mostly as a by-product of unrelated programs and facilities rather than by special design. We do, however, for the first time, have a direct, *in situ*, collection and identification of stratospheric particles undoubtedly connected to the volcanic event [Mossop, 1964].

Mount Agung, a dormant volcano on the southeast shore of Bali, at about 8°25'S 115°30'E, "blew its top" on 17 March 1963, resulting in considerable loss of life and damage in its immediate vicinity [Booth, Matthews, and Sisson, 1963]. Among the first to draw attention in print to subsequent skylight effects, at least in the northern hemisphere, were the Meinels [Meinel and Meinel, 1963]. In their short note, they not only related the unusual twilights seen from Tucson, Arizona, with the incursion of Agung's dust but also estimated the height of the main stratum as 22 km from simple considerations of the earth's shadow.

The unusual duration and delicate coloration of the evening twilights, also quite noticeable locally (Southern California coast) during the fall and winter of 1963-1964, were probably not very different from the Krakatoa twilights. Under very clear sky conditions produced by so called "Santa Ana" weather (dry, off-shore local winds) even a secondary glow, similar to the Krakatoa phenomenon mentioned in Sec. 2.1.3, was often observed by the writer. I also noticed both the asymmetry mentioned by the Meinels [1963]--i.e., the fact that the evening twilight glows, during the above period, were centered some 10° to 20° south of the (sub-horizon) sun's azimuth--as well as nonuniformities in the glow, clearly suggesting "patchiness" in the dust stratum. These characteristics are also mentioned by Volz [1964], in his observations near Stuttgart, who thus appears to be the first to relate northwestern European twilights with the Agung volcanic dust.

As to estimates of the initial amount of solid material injected into the atmosphere by this volcano (such as were given for Krakatoa and Katmai), none seem to have been published to our knowledge. One therefore has to rely on whatever turbidity data are available to arrive at such an estimate indirectly, if at all, by comparison with turbidity produced by the previous volcanic veils in relation to the magnitude of the event.

4.2 The Agung Turbidity

That the Agung eruption of 1963 resulted in a considerable increase of turbidity, at least in the stratosphere, also follows from the unusual darkness of the totally eclipsed lunar disc observed on 30 December 1963 [Brooks, 1964]. In his article, Brooks lists (after F. Link) previous dark lunar eclipses, most of which are clearly connected with major volcanic events such as Krakatoa (cf. Sec. 2.1.1) and Katmai. Interestingly, the degree of darkness is not always proportioned to the estimated amount of volcanic material injected into the atmosphere. In general, this phenomenon cannot be used as a measure of the overall increase in turbidity (nor for estimates of changes in the planetary albedo) since it can be shown that small changes in *stratospheric* turbidity will cause large changes in the illumination of the eclipsed moon [Brooks, 1964].

In principle, the turbidity anomaly from volcanic dust could be estimated from the analysis of rather complete data on the brightness, spectrum, and polarization of the twilight sky--*provided* there exist complete solutions of the corresponding mathematical problem for realistic models of the spherical, inhomogeneous atmosphere. This is not the case; and, in general, the twilight method at present should be considered as the *least* reliable one for quantitative estimates here. For this reason, we cannot rely overly on Volz's [1964] derivation of the vertical profile of the atmospheric turbidity for the winter of 1963-1964 in estimating the Agung turbidity anomaly. In fact, as becomes clear from a later elaboration by the same author [Volz, 1970a], his deductions are based merely on the analysis of brightness ratios at a fixed point in the sky as a function of solar-depression angle. The underlying theory and the reliability of results

obtained thereby have been seriously questioned [cf. Rozenberg, 1963]. Volz [1970a], in his own paper, which contains a number of specious and unsubstantiated opinions, seems to agree when he concludes that his "simple twilight method seems well suited (for monitoring the stratospheric dust) *until reliable, more sophisticated methods*" become available (italics mine). However, this is not to deny the value of this particular technique in the *detection* of stratospheric aerosol layers [Shah, 1969].

At least two other methods do exist, which are rather more reliable—though less sophisticated--than any twilight method, since they both are based on fewer assumptions and a simpler theory. As we have seen, one is the old Smithsonian "long method," which uses the sun as a source, yielding turbidity values for the entire atmospheric layer. The other, based on recent technology, uses the pulsed-laser technique and is capable of indicating the vertical profile of the turbidity [Grams and Fiocco, 1967].

To our knowledge there exist no atmospheric transmission data from solar work, similar to those after the Katmai eruption, that could be used in the case of the Agung 1963 event. As mentioned above, no special observing programs were organized for this purpose; but some useful information is provided by stellar extinction data. We attempt to use these to obtain some idea about the nature of the Agung turbidity, by a reduction similar to that summarized in Fig. 1 for Katmai.

The most detailed data seem to be those presented by Irvine and Peterson [1970], particularly those from Boyden Observatory located in South Africa at about 29°15'S 26°E. These authors list stellar extinction coefficients for various wavelengths, obtained by filters, in "magnitudes per unit air mass." To convert these traditional and awkward units (so dear to the hearts of astronomers) to the more rational negative exponent called *optical thickness*, one has to multiply their listed values by

$$100^{-\frac{1}{5}} \log_e 10 = 0.917.$$

We have used this procedure on selected data taken on several nights during 1963 and 1964, presumably reflecting turbidity conditions during and after the passage of Agung dust over the observatory. The results are plotted on Fig. 2. Here again the straight line marked $\tau_{1R}(1350)$ represents the Rayleigh optical thickness over Boyden Observatory, whose height is given as 1387 m. The other labeled curves (formed by straight lines joining the datum wavelengths) were obtained as follows from Irvine and Peterson's [1970] original data:

Curve (a): Average optical thickness for seven "good" nights of May 1965 (consistently low extinction period) less the Rayleigh value.

Curve (b): Average for five consecutive "fair" nights in August 1963 less the Rayleigh value. The increased extinction observed during this period may be attributed to Agung's dust, according to the authors.* The dashed portion of this curve is to indicate the absence of data at $\lambda 0.634\mu m$, near the Chappuis ozone absorption maximum.

Curve (c): Observed values on night of 10 September 1963, marked "good," less the Rayleigh value; apparently the highest extinction observed in the blue to green filters among the data given.

Curve (d): Obtained by simply subtracting the unprocessed data corresponding to curve (a) from those for curve (b) and converting to optical thickness, $\tau_D(\lambda)$.

In our analysis we chose the May 1965 values as a standard for clear nights, assuming that the Agung dust effects had practically disappeared at Boyden by that time (data for only two pre-Agung nights are shown by Irvine and Peterson). Thus curve (a) in Fig. 2 is the

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*The erroneous reference to a volcano in northern Australia [Irvine and Peterson, 1970, p. 66] should be corrected to read "Agung volcano in Indonesia," according to a private communication by F. Peterson, December 1970.

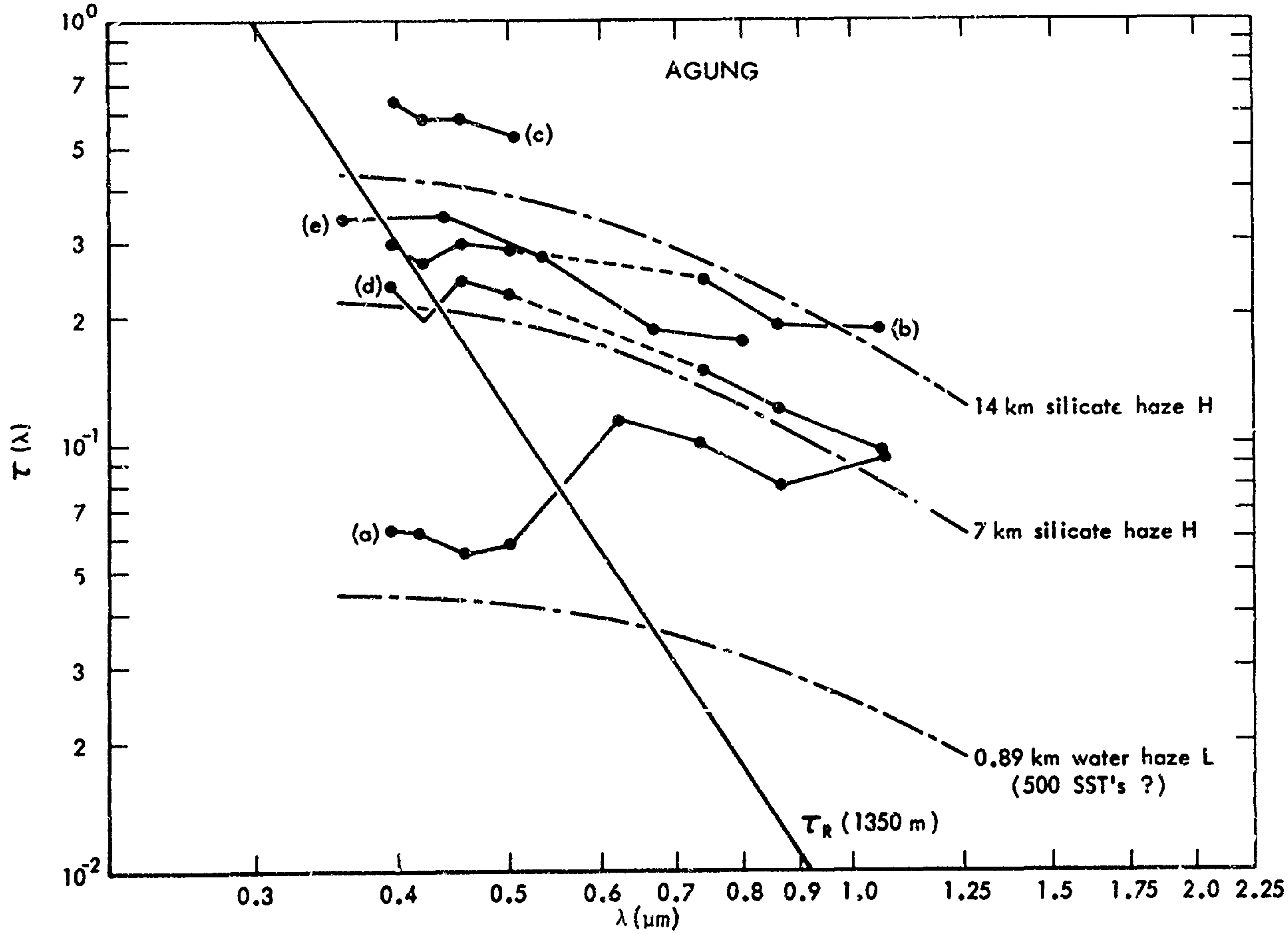


Fig. 2 -- The turbidity anomaly produced by the Agung (1963) eruption, evaluated on the basis of stellar extinction values obtained from a South African and Austrialian observatory (see text).

is the equivalent of curves (a) and (b) of Fig. 1 (p. 21) for clear days at Mt. Wilson. There seems to be little resemblance in the shape of these sets of curves (there is no reason why there should be any) except for the ozone absorption feature at $\lambda 0.6\mu m$. However, after processing the data as in curve (d), a smoother wavelength dependence is obtained with negative slope, similar to the Katmai data. We may reasonably assume that the processing has eliminated the ozone absorption in the $0.50 < \lambda < 0.73\mu m$ interval and that the dashed line in curve (d) joining these two points is a good interpolation, especially since it has almost the same slope as that of neighboring portions on either side. (We have no explanation for the dip between $\lambda 0.40$ and $0.45\mu m$ appearing in both curves (b) and (d) at this time, and assume that it is not related to the volcanic dust component.)

A search for an independent set of data of this type on Agung turbidity was largely unsuccessful. Przybylski's [1964] description of some stellar extinction measurements from a South Australian site during 1963 are of little help here because there is no wavelength resolution given and the analysis is not clear. He does mention that the extinciton in "visual light," during the August 1963 peak turbidity period, reached a value three times the pre-Agung norm, which seems to be borne out by the ratio between curves (b) and (a) in Fig. 2.

The earlier short note by Hogg [1963] of the same observatory (Mt. Stromlo) is much more informative and to the point. He reports observations of an unusually large and persistent "meteorological corona" around the sun after Agung, which we interpret as an aureole (rather than a Bishop's ring, as interpreted by Volz [1970b], who mentions two further unpublished ring sightings by observers other than himself), in the absence of any mention of colored rings of the proper radius (see Sec. 2.1.2). Hogg also quotes stellar extinction values (in magnitudes) at five wavelengths obtained variously in June and July 1963, together with a set of "normal" values at the same wavelengths. Curve (e) in Fig. 2 represents the difference of these two sets without further processing except for the conversion to optical thickness mentioned before. As may be seen, the Australian data, all things considered, are remarkably similar to the South African ones, represented in curve (d), both in magnitude and slope.

Assuming that our above analysis of the data is valid--and some of these *do* lend themselves to other methods of analysis and interpretation--we may compare the curves in Fig. 2 directly with those of Fig. 1 to draw some inferences. Clearly, the Agung dust extinction curves, $\tau_D(\lambda)$, although of the same magnitude as those for Katmai, show a steeper slope than the latter, at least in the available range $0.4 < \lambda < 1.0\mu m$. This indicates that there must have been a smaller proportion of large particles in the Agung dust than in that of Katmai (and most probably of Krakatoa). Thus we find that the best choice for Agung is a "haze H" model [ESSP, p. 78 and T.102, T.103], which follows a size-distribution law of the form

$$n(r) \propto r^2 \exp\{-20r\} \ . \tag{5}$$

As may be seen from Fig. 3, where we plot the logarithm of the size distribution function, n(r) (normalized to 100 $cm^{-3}$), as a function of the radius r, the H model has a significantly larger concentration of small particles in the range $0.06 \leq r \leq 0.25\mu m$ than the L model; but for $r > 0.25\mu m$, the concentration in the latter model considerably exceeds that in model H. The verification of either model, by the way, would entirely preclude the formation of any Bishop's or other corona-type rings around the sun.

The equivalent depths of homogeneous model H haze needed to bracket the Agung data are 7 and 14 km, respectively, as shown by the dot-dash curves correspondingly labeled in Fig. 2. The designation "silicate" merely indicates a dielectric substance with real index of refraction between 1.54 and 1.56 in the visible range. Water substance with the same size distribution of particles would result in a steeper slope than given by the data. Also, the layer thickness shown should not be taken literally but merely to indicate the total amount of material needed to produce the required turbidity regardless of concentration.

In this case, taking the upper figure of 14-km haze H and proceeding as before [ESSP, p. 78], we get

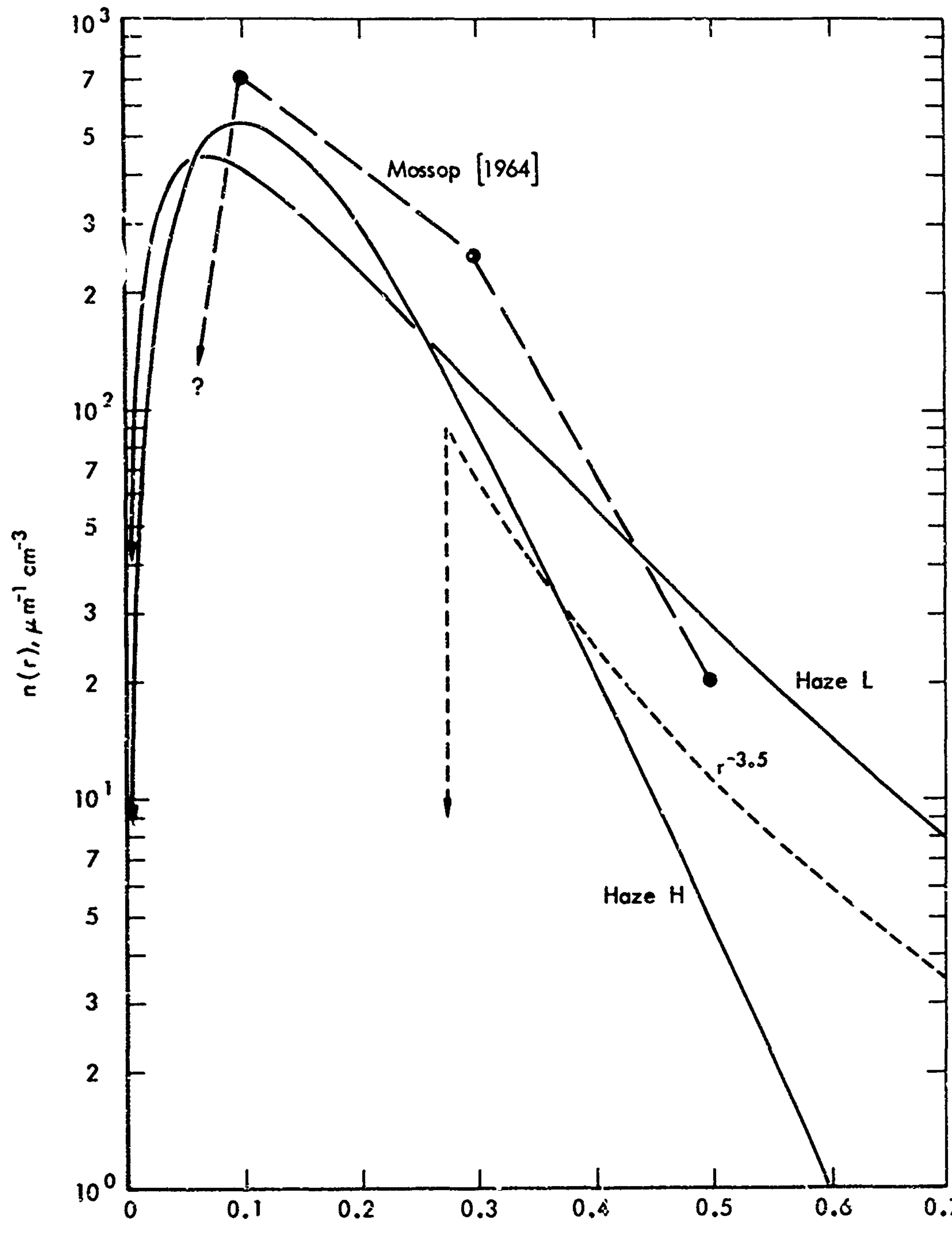


Fig. 3 -- Comparison of proposed volcanic dust size-distribution models (full curves) with actual stratospheric sampling of Agung's dust (full dots) and the distribution proposed by Grams and Fiocco (dotted curve).

$$14 \cdot 10^5 \cdot 3.142 \cdot 10^{-12} = 4.4 \cdot 10^{-6} cm^3 (cm^2 column)^{-1}$$

for the bulk volume of the dust. (This is less than one-half of the comparable figure for Katmai or $1.05 \cdot 10^{-5} cm^3$ per $cm^2$column). Assuming that the above amount of Agung dust was distributed over the same intertropical zone as we used for Krakatoa, we arrive at a total of

$$4.4 \cdot 10^{-6} \cdot 2.04 \cdot 10^{18} = 9 \cdot 10^{12} cm^3$$

of dust material. From this point of view, and comparing this figure with similar ones we estimated for the other volcanoes, the Agung eruption of 1963 injected an amount of dust equivalent to 70 percent of that injected by Katmai in 1912 and 31 percent of that injected by Krakatoa.

Finally, on the basis of the ratio 1/600 deduced earlier for Katmai and Krakatoa, the total volume of ejecta from Agung would be

$$600 \cdot 9 \cdot 10^{12} = 5.4 \cdot 10^{15} cm^3$$

or 5.4 $km^3$ of material, which of course is in the same ratio as above to the total volumes ejected from the other two, and assumes the same type of volcanic event. It would be interesting to see whether independent volcanological surveys arrive at a comparable estimate for Agung.

Note that curve (c) in Fig. 2, which represents the highest non-Rayleigh extinction observed at Boyden Observatory [Irvine and Petersen, 1970], covers too small a wavelength range for detailed analysis. Its magnitude of about ten times the clear-day values in curve (a) is remarkably close to the value of 0.63 we estimated for Krakatoa's initial turbidity at $\lambda 0.5 \mu m$. However, as neither blue and green suns nor Bishop's rings were definitely reported (to our knowledge) in the case of Agung, we conclude that the latter's dust did not contain nearly as many large particles as Krakatoa's but conformed to a different size-distribution law.

Turning to the optical radar method, Grams and Fiocco [1967], after a clear and concise review of recent work on stratospheric aerosols, present and discuss their own use of the pulsed laser to determine stratospheric turbidity during 1964 and 1965. In particular, after corroborating the existence of a stratospheric aerosol maximum near 20 km as a permanent feature between about 60°S and 70°N latitudes [Junge and Manson, 1961], they looked for anomalous increases in the northern hemisphere that might be attributed to Agung's dust.

Grams and Fiocco [1967] used the well known pulsed, ruby laser system ($\lambda 0.6943\mu m$), which is capable of a spatial resolution of 0.015 km and reduces to an effective resolution between 0.5 and 1.0 km after the necessary smoothing of the raw data. In an attempt to minimize the ambiguities inherent in the interpretation of such data [Deirmendjian, 1965; ESSP, p. 124], these authors used the ratio of the actual signal to a hypothetical norm, based on the return from the region of minimum turbidity between 25 and 30 km as an index of the turbidity and the exponential structure of the atmosphere. This method is attractive in that it involves a minimum of assumptions and relies on a more realistic background than a theoretically derived one.

The average of all the ratios thus obtained by Grams and Fiocco [1967] on a number of nights during 1964 and 1965 from Lexington, Massachusetts, and College, Alaska, was found to vary between 1.5 and 2.3 at its maximum, centered near an altitude of 16 km. On the basis of the often used but awkward Junge size-distribution model, and assuming Mie scattering with refractive index 1.5, these authors then deduce a local extinction coefficient at 16 km equivalent to $2 \cdot 10^{-3}$ $km^{-1}$ and a particle concentration of 0.9 $cm^{-3}$. From this and their average profiles between 12 and 24 km, they arrive at a total of $6.8 \cdot 10^{5}$ particles per $cm^{2}$ column in that layer equivalent to an optical thickness $\tau_D = 0.015$ presumably at the laser wavelength near $\lambda 0.7\mu m$. They also give a mass of $6 \cdot 10^{-7}$ g per $cm^{2}$ column for particles of an assumed density 2, equivalent to a bulk volume of $3 \cdot 10^{-7}$ $cm^{3}$ per $cm^{2}$ column in the units used here. This may be compared with the 2.2 to $4.4 \cdot 10^{-6}$ units we estimated above for Agung's dust on the basis of the stellar extinction data.

Thus Grams and Fiocco's [1967] own analysis of their laser data from northern latitudes yields about 1/10 of the bulk amount of Agung related aerosols we independently deduced for southern latitudes from a different set of data. This is consistent with their estimate of 0.015 for the optical thickness of the dust layer compared with values exceeding 0.15 near λ0.7μm we deduced as shown in Fig. 2. We note that recently Volz [1970b], in a somewhat confusing article, after considering some of the same stellar extinction material mentioned here, together with a few of his own (?) turbidity measurements at λ0.50μm by unidentified methods, concludes that the Agung dust hardly affected the normal "background" values of northern hemisphere (32° to 47°N) attenuation after 1965.

In fact, quite likely the laser-derived amount may have been *overestimated*, considering the ambiguity inherent in the reduction and analysis of monostatic laser-radar data, and the particular size-distribution model used in its derivation [Grams and Fiocco, 1967], i.e., a power-law dependence in the form

$$n(r) \propto r^{-3.5}, \qquad 0.275 \leqq r \leqq 3.3\mu m. \qquad (6)$$

The shape of such a function (multiplied by an arbitrary constant) is indicated by the dotted curve in Fig. 3 for comparison with the continuous distribution models we use here in the interpretation of the Katmai and Agung extinction data. The distribution (6) entirely omits the smaller particles in the range $0.01 \leqq r \leqq 0.275\mu m$ where the concentration of stratospheric particles should be high [Junge and Manson, 1961]. At the same time, a distribution such as (6), when integrated with respect to the Mie extinction coefficient, shows an increasing contribution from the larger particles depending on the upper limit [ESSP, p. 80] (3.3μm in this case, outside the frame of Fig. 3, where the slope of the dotted curve becomes much flatter than that of either curve L or H). Furthermore, (6) seems to disagree with accepted size ranges and distributions for stratospheric aerosols, and even with actual counts of Agung particles over Australia (indicated by the dashed trace in Fig. 3) a year after their injection [Mossop, 1964].

At any rate, the theory of polydisperse Mie scattering shows that the mass extinction coefficient of a polydispersion increases with increasing proportion of smaller particles in the distribution [ESSP; cf. parameters on p. 78 and coefficients in Part II tables]. Hence, quite possibly the laser data could be interpreted in terms of a size-distribution model with smaller particles, say within the range $0.10 \leq r \leq 1.5\mu m$, even with the use of a power-law function as in (6), in which case Grams and Fiocco's [1967] mass estimates would be reduced by a factor of 2 or 3. This is further reinforced by the fact that, for a given mass, the smaller particles in a polydispersion contribute most to the back-scattering cross section [ESSP, p. 90].

Finally, we may consider Mossop's [1964] actual sampling of the Agung dust from a U-2 aircraft before and after the eruption. At 20 km over Australia, he did find a noticeable increase in the concentration (by a factor of 3), as well as in the average size of the irregular particles he collected. Their size tended to diminish again with the passage of time. On Fig. 3 we show only one of Mossop's distributions (multiplied by an arbitrary constant), as derived from his counts and microscopic measurements, collected between 15° and 35° south latitude on 2 April 1964, a year after the Agung event. Although the resolution given is very gross--here plotted every 0.2μm in the "radius" deduced from Mossop's "diameter"--the maximum concentration around 0.1μm and the slope in the $0.3 \leq r \leq 0.5\mu m$ are rather faithfully reproduced by our model H distribution. The latter, as mentioned above, provided a very good interpretation of stellar extinction at Boyden Observatory, South Africa, in August 1963, as shown in curve (d) of Fig. 2. If we accept Mossop's [1964] deductions regarding the time variation of the dominant size and size range of Agung particles, we note that his data for the same period indicate greater values than our model. We must conclude that if we indeed are dealing with Agung particles in both cases, the dust layer must have lost the larger end of the size distribution somewhere between Australia and South Africa--a reasonable deduction.

In general, it appears that the optical-radar method of sounding the stratosphere for its aerosol content can provide useful information

when properly used. In this case, it is regrettable that the method was not used simultaneously and at the same location where the Agung dust effect was most evident on the stellar extinction in the southern hemisphere.

### 4.3 Other Agung-Related Effects

One of the most interesting aspects of the Agung event--indeed of any volcanic activity that affects the optical weather--is the rate of spreading and the extent of its dust veil as a function of geographic location and season, and elapsed time from the main eruption. Here again, despite the obvious importance of such information to the meteorological sciences, no coordinated observational program was organized after the magnitude of the explosion became known. The best attempt in this direction seems to be that of Dyer and Hicks [1965, 1968]. This effort is mainly based on a rather crude dust index that they define in terms of the reduction of the (unresolved) direct solar flux, as measured routinely at a number of meteorological stations over the globe. Despite the gross simplifications involved in their analysis and the sparsity of data, these authors have been able to present a coherent and plausible picture of the spreading of Agung's dust. The most reliable and interesting of their conclusions seem to be: (a) that the initial injection height was about 22 to 23 km, creating an equatorial reservoir of stratospheric dust; (b) that most--but not all--of this dust remained in the southern hemisphere with a gradual decrease in total amount over subsequent years; (c) that there was a winter dust maximum in each hemisphere, uniformly in phase between 30° and 90° latitude, with an apparent poleward progression with time; (d) that, if a choice must be made, the main spreading agent may have been the mean large-scale motion of the atmosphere rather than eddy diffusion.

The last two inferences are strongly dependent on the reliability of the method, as pointed out by Clemesha [1971]. The latter's critique, however, stems from an unwarranted reliance on the analysis of the laser-radar data, as pointed out by Dyer [1971] in his reply and by this author some time ago [Deirmendjian, 1965].

Another interesting phenomenon, which may be indirectly related to Agung, was suggested by Pittock [1966]. He detected a rather sharply defined ozone-deficient layer between 20 and 21 km over Boulder, Colorado, in the routine ozonesonde records for March/April 1964. He then observed that this coincided with an aerosol layer at the same height whose existence was apparently deduced from a single observation of abnormal attenuation of $\lambda 0.44 \mu m$ light from the setting sun, as reflected on a large meteorological balloon floating at 33 km. Although the reliability of the dust detection may be questioned on several grounds, Pittock's suggestion that it originated over Agung and that the entire layer was transported by horizontal winds to Boulder without much change is intriguing. Interestingly, the author does not attribute the ozone deficiency to the local action of the volcanic particles themselves, but rather to a conservation of the initial properties of the tropical stratospheric air found over Boulder.

These latter conclusions were essentially confirmed by Grams and Fiocco [1967] who, from an independent analysis of ozone soundings over Bedford, Massachusetts, found a statistically significant anti-correlation between total ozone and stratospheric dust amounts. These authors also refrain from attributing the above to a direct causal connection between the presence of dust and the reduction of ozone amounts.

Another possible effect has been suggested by Newell [1970a,b] who, in a pair of short notes, attributes a 5° positive anomaly in stratospheric temperatures, observed over northwest Australia late in 1963, directly to a supposed absorption of sunlight by Agung particles. Later Sparrow [1971] pointed out that the anomaly might be just as plausibly attributed to a breakdown of the so-called quasi-biennial oscillation found in recent meteorological soundings over the tropics. In any case, the effect of absorbing and emitting aerosols on the ambient atmospheric temperature is not well understood nor experimentally verified. Until the possible existence of such an effect can be proven, its use to explain certain temperature anomalies, before considering other likely possibilities, can hardly be justified.

Finally, we might mention the unique--by location--measurements of direct sunlight and global radiation at the South Pole (about 3200 m above mean sea level) described by Viebrock and Flowers [1968]. In particular, their data for normally incident direct sunlight measured with an Eppley pyrheliometer on cloudless days clearly show the effect of an attenuating layer after November 1963 (data for September and October 1963 were unavailable). It is unfortunate that additional measurements in narrow bands were not conducted simultaneously to provide information on the wavelength dependence of the turbidity. This would enable us to compare the antarctic data with those discussed in Sec. 4.2 and to consider the likelihood of Agung dust as the attenuating layer over the South Pole, as suggested by these authors. As it is, from their tabulated data for unresolved radiation on 10 February 1964, a rather turbid day [Viebrock and Flowers, 1968, Table 3], one may deduce an "average" optical thickness, $\bar{\tau}_D$, over the spectral range of the instrument, by equating air mass with the secant of the solar zenith distance and by putting 0.68 (due to the elevation) for the normal air mass. We thus have

$$\frac{2.054 - 0.884}{2.054} = 0.570 \equiv \exp\{-0.68\bar{\tau}_D \sec 75°26'\}$$

where 2.054 is the reported extraterrestrial energy and 0.884 is the amount depleted by aerosol scattering, whence

$$2.70\bar{\tau}_D = 0.562$$

yielding a value of $\bar{\tau}_D = 0.208$ for the average optical thickness of the dust layer. The same type of reduction applied to the 72.3 percent of "normal" intensity, observed on 18 December 1963 (also a highly turbid day), yields $\bar{\tau}_D = 0.191$. Comparing these values with the estimates shown in Fig. 2 for Agung's dust turbidity anomaly, $\tau_D(\lambda)$, we see that they fall well within the values for the middle of the visible range, obtained from data closer to the volcanic source in latitude and time. On this basis, then, it is not at all implausible

to assume that the anomalous attenuation of direct sunlight, reported over the South Pole in late 1963 and early 1964, was entirely due to Agung's dust layer, transported essentially unmodified--whether by advection or otherwise--from the source region.

Viebrock and Flowers [1968] also list values of the global radiation, G, or the sum of direct sunlight, S, and entire sky radiation, H, incident on a *horizontal* surface, at Amundsen-Scott Station. These are given in terms of the ratio of the measured flux to the equivalent theoretical value $G_R$ for a molecular horizontal atmosphere over a Lambert surface with 0.80 reflectivity, determined some years ago [Deirmendjian and Sekera, 1954]. The most interesting result here is the rather high values of the ratio $G/G_R$--0.89 to 0.97--observed during the antarctic summer months of 1961 and 1962, despite the simplifying assumptions we had used in our model [Deirmendjian and Sekera, 1954]. However, as pointed out in that early study, changes in the global radiation are not reliable indicators of the presence, amount, or type of turbidity, simply because such radiation represents an integral over too many variables that may be mutually compensating. Thus, it is not surprising that, after the presumed onset of Agung dust over the South Pole, there was only a modest decrease in global radiation, a decrease the cause of which it is difficult to determine among various likely candidates.

On the other hand, it is well known that for cloudless skies and under moderate turbidity conditions an increase in turbidity (due to predominantly *scattering* aerosols) will result in a lowering of S together with an increase in H. Therefore, the ratio S/H will drop even faster with increasing turbidity, an effect that is intensified over a highly reflecting surface and at low solar elevations [Deirmendjian and Sekera, 1954]. Both these conditions, of course, are precisely those found in Antarctica, and the ratio S/H, easily obtainable by means of the instrumentation used there, would be a very sensitive general index of turbidity. Nevertheless, Viebrock and Flowers [1968, Table 1] do not tabulate and analyze this parameter in detail, preferring the somewhat less sensitive ratio H/(S + H) as a turbidity index. From their limited published data, we note that the value of

S/H was as low as 0.875 in February 1964 compared to a high of 4.42 in February 1960. These may be compared with S/H values we had derived from old Smithsonian Observatory data [Deirmendjian and Sekera, 1954, Fig. 4]. The high South Pole value mentioned above falls well within the 1917 observations (for a solar zenith distance 75°.5) at Hump Mountain, N. Carolina, under clear conditions. The low South Pole value of 0.875 is well below that of about 2.0, observed at Mt. Wilson, California, in September 1913, when the effects of Katmai's dust were presumably still present. If we assume that this is a true difference--and not the result of instrumental and other extraneous discrepancies--we must conclude that the turbidity over the South Pole in February 1964 was higher than that over Mt. Wilson in 1913, even after allowing for the height difference in the two stations. Such a conclusion, however, is at variance with our own derivation of the overall magnitude of the turbidity introduced by Katmai and Agung, respectively, on the basis of the more direct evidence provided by attenuation of sunlight and starlight. If corroborated, therefore, the high concentration of Agung dust over the South Pole must be explained in terms of atmospheric transport mechanisms capable of producing such concentrations.

## 5. CLIMATIC EFFECTS OF VOLCANIC DUST

In this section we shall briefly examine weather and climate changes--in the sense of the more traditional meteorological parameters--proposed in the literature as possible results of the injection of volcanic dust into large portions of the atmosphere. In a recent monograph, H. H. Lamb [1970]* has undertaken a detailed review and discussion of this subject. In particular that author, after collecting and tabulating all known volcanic activity on the basis of available records, has examined them with a view to their classification--for the first time--in terms of a quantitative "dust veil index (d.v.i.)." The underlying idea is presumably a better understanding of possible climatological effects in terms of the magnitude of such an index--a reasonable and potentially useful exercise.

Upon closer examination, however, this attempt fails in its purpose essentially because of the author's failure to provide a clear and precise definition, not only of the index itself but also of the parameters involved, and of his rather loose discussion of various related meteorological concepts and processes. This is self-defeating, even granting the existing fragmentary and qualitative description of the effects of all but the most recent volcanic events. For example, at the very outset [HHL, p. 471] the "d.v.i." is related to the "total loss of incoming radiation . . . occasioned by each eruption," i.e., the reduction in the *global* radiation--whereas what the author seems to have in mind is simply the reduction in the *direct* sunlight.

In fact three separate, purely *empirical* formulas are used to determine the "d.v.i.," each based on a different criterion, and each used arbitrarily merely depending on the *availability of information* in each case: reduction in insolation, reduction in "mean temperature," and volume of dust material, respectively, all weighted by the extent and duration of the "dust veil." In view of the unreliability of most of the source material and the diversity of these criteria, the "d.v.i." numbers thus obtained can hardly be expected to be mutually

*Hereinafter referred to as "HHL."

comparable and uniformly significant. Whereas two of the above criteria, as we have seen, may be related to the magnitude of the dust effect, a supposed reduction in "mean (surface) temperature," despite the author's unsubstantiated claims [HHL, pp. 460-469] may by no means be related to the presence of volcanic dust until considerably more evidence than we now possess becomes available. Potentially the most meaningful "d.v.i." criterion, the optical thickness anomaly, $\tau_D(\lambda)$--depending as it does on the amount, type, and size distribution of the volcanic dust particles--is either not mentioned or is glossed over in the most simplistic terms (HHL, pp. 460-461; *vide* his unjustified use of the geometric optics cross section based on the authority of Humphries [1940]).

Nevertheless, for completeness, we compare below Lamb's "d.v.i." numbers for the three volcanic events we have been discussing with our own estimates of relative amounts of injected dust, adjusted to the same scale of 1000 units for Krakatoa employed by Lamb:

| | Krakatoa | Katmai | Agung |
|---|---|---|---|
| d.v.i. (Lamb): | 1000 | 500 | 800 |
| This work: | 1000 | 443 | 310 |

The values in the last line are adapted from the ratios we arrived at in Sec. 4.2 on the basis of the optical thickness anomaly, definite aerosol models, and an assumed spread area for the dust, without having in mind any intended use as a volcanic dust index. Comparison of the two evaluations shows that, whereas they agree for Katmai, Lamb's "d.v.i." for Agung is considerably overestimated (by a factor of 2.5) from our point of view.

Because of our reservations mentioned above, and Lamb's frequent indulgence in conjecture based on specious arguments, we are inclined to discount most of his other conclusions [HHL, p. 453] without further analysis and discussion here. Undoubtedly, provided one does not take the "d.v.i." evaluations too seriously, his detailed and annotated tabulation [HHL, App. I] of most known volcanic events in chronological order should be useful to the interested student of

climatology. (We note that the important explosions of Mexico's Paricutín volcano in 1943 and 1945 are omitted from Lamb's list.)

Another recent report [SCEP, 1970], prepared by an *ad hoc* summer study group on problems of man-made modification of the global environment, deserves some mention here. Of particular interest is the group's attempt to evaluate the climatic effects of the particulate loading of the lower stratosphere as a result of future commercial flights of supersonic transports (SSTs) on a world-wide scale. Such effects, if any, would mainly arise through the scattering and absorption properties of the particles and their effect on the radiation balance of the stratosphere and perhaps the earth's surface. It is well known that present knowledge in these areas is incomplete and subject to speculation. However, very few names of scientists of known competence in these fields appear in the roster of full- and part-time participants in the SCEP's working sessions, despite the implication contained in the report's introduction that "some of the world's leading scientists" were participating [SCEP, 1970, p. 6]. The report's conclusions, therefore--at least on this subject--may not carry much weight as they seem to be based on second- and third-hand information and conjecture rather than any substantial scientific evidence and discussion such as might have been introduced by the (absent) active workers themselves. Nevertheless, since these conclusions were mostly arrived at on the basis of Agung's dust and its presumed meteorological effects, we shall briefly compare the SCEP's estimates of SST particulates with our present estimates of volcanic dustiness.

The amount of particulates was based on the exhaust products of an estimated 500 SST craft operating at about the 20-km level for 2500 hours yearly each [SCEP, 1970, pp. 71-74]. If we take the "peak N. hemisphere" value of 3.46 parts per billion (by mass), as given by the report, we get a total of

$$3.46 \cdot 10^{-9} \cdot 3.85 \cdot 10^{20} = 1.33 \cdot 10^{12} \text{ grams}$$

for the *SST particulate loading*, assuming a two-year residence, where $3.85 \cdot 10^{20}$ is the mass of one-half of the stratosphere, as given in the

report. This may be considered as a *worst-case estimate*, since the above peak value was taken as ten times the estimated global average. Assuming further that the SST particulates have unit density (again a worst case) this represents condensed particulate material with a total volume of $1.33 \cdot 10^{12}$ cm$^3$. This means that the SSTs *at most* could add a particulate load equivalent to some 15 percent of that estimated in Sec. 4.2 for the Agung volcano ($9 \cdot 10^{12}$ cm$^3$), or 47 of Lamb's "d.v.i." units (see above).

If, for the sake of comparison, we assume that the SST particulates will be confined to the area above 30°N or one-fourth of the earth's surface, we have

$$\frac{1.33 \cdot 10^{12}}{1.275 \cdot 10^{18}} = 1.04 \cdot 10^{-6} \text{ cm}^3 \text{ (cm}^2\text{, column)}^{-1}$$

of equivalent condensed material. Proceeding as before in terms of our polydisperse models mentioned in Secs. 3.2 and 4.2 and shown in Fig. 3 [ESSP, pp. 78, 166, 177], this amounts to 0.89 km of water haze L resulting in a turbidity anomaly, $\tau_D(\lambda 0.45) = 0.043$; or 3.32 km of water haze H with $\tau_D(\lambda 0.45) = 0.068$. In either case, *the expected turbidity anomaly from future SST flights, at its worst, will hardly exceed the background turbidity found above mountains on clear days away from city pollution.* This is rather less than the report's estimate of SST particulate loading [SCEP, 1970, p. 16] either on the global scale or at peak concentration.

In view of this conclusion, the other projections about possible climatological disturbances directly attributable to SST particulates, mentioned in the SCEP report (p. 16), cannot be considered seriously insofar as they are based on presumed Agung dust effects. By this, we do not mean to imply that the constant injection of SST (or similar) exhaust products into the lower stratosphere presents no environmental problem nor, in general, that the concerns reflected in the SCEP report are unjustified. We merely refer to its treatment of the particulates problem as it relates directly to the present study.

Similar remarks apply to the report's conclusions [SCEP, 1970, p. 61] on the man-made component of tropospheric turbidity and projections thereof, since here too the "substantial evidence" for a dangerous global trend turns out to be based on rather shaky data not corroborated by other recent work [Ellis and Pueschel, 1971; Twomey, 1971]. In this respect, we tend to agree with the general evaluation presented in H. Landsberg's [1970] excellent article: that whereas man-made pollution is quite capable of affecting the local climate near major sources, on the global scale such effects have neither been positively observed nor yet proven to be possible theoretically.

## 6. SUMMARY AND CONCLUSIONS

In the light of increasing attention being given recently by scientists and others to the presumed role of particulate turbidity in past and future climatic changes, we have here undertaken a critical review of the subject in order to judge the validity of such claims. We have endeavored to separate fact from conjecture and speculation by choosing for particular study and intercomparison the three major volcanic events, known to have introduced considerable amounts of particulates, that have been best documented: Krakatoa (1883), Katmai (1912), and Agung (1963).

Our survey shows that, whereas all three eruptions resulted in clearly recognizable turbidity anomalies over large areas of the earth for periods of a few years, little solid evidence exists of climatological (or weather) effects in terms of anomalies in the conventional meteorological parameters. In comparing the volcanically produced turbidities with that to be expected from the operation of future commercial supersonic transports (SSTs), we find that the latter would be but a fraction of the turbidity introduced by a single "low yield" volcanic event such as Agung's.

In more specific terms and with the understanding that our Krakatoa evaluations are arrived at indirectly on the basis of data from the two more recent events (with our "worst case" SST evaluations also shown in parentheses for comparison), our principal findings may be summarized as follows:

(i) Typical turbidity anomalies, in terms of absolute increments in optical thickness in the visible region ($\lambda = 0.5\mu m$), for the particulate layer as observed away from urban pollution sources, say within a few months after the volcanic events, may be given as: Krakatoa, 0.55; Katmai, 0.35; Agung, 0.25; (500 SSTs, 0.05).

(ii) Mainly on the basis of the above optical criterion and additional considerations of the nature and extent of the disturbances, the overall magnitude of the turbidity an ..lies

produced in each case may be rated as follows, in relative units: Krakatoa, 1.00; Katmai, 0.44; Agung, 0.31; (500 SSTs, 0.047).

(iii) The physical and chemical characteristics of the volcanic particles have not, so far, been completely and accurately determined for any of the three events. However, well-defined polydisperse (optical) scattering models may be easily fitted to the observed, wavelength-dependent turbidity. These indicate that the particles may have been composed of non-absorbing (or very weakly absorbing) dielectric material with refractive index close to that of water (1.33) or silicate (1.55). Whereas the shape of the particles may have been amorphous rather than spherical or crystalline, their size and size distribution appears to conform to that of other natural aerosols normally found in the upper troposphere and lower stratosphere.

Assuming a solid or liquid substance of unit density, typical local masses of the volcanic dust content for conditions as in (i) may be estimated as: Krakatoa, $1.5 \cdot 10^{-5}$; Katmai, $9.0 \cdot 10^{-6}$; Agung, $2.9 \cdot 10^{-6}$; (500 SSTs, $1.0 \cdot 10^{-6}$) gram per $cm^2$ column, respectively. (These figures are not exactly in the same ratio as the turbidity estimates in (i) as they depend on the scattering model and the spread of the layer assumed in each case.) Likely values for the total mass of material (of unit density) in each case may be: Krakatoa, $3 \cdot 10^{13}$ g; Katmai, $1.34 \cdot 10^{13}$ g; Agung, $9 \cdot 10^{12}$ g; (500 SSTs, $1.33 \cdot 10^{12}$ g). The most massive injection, that of Krakatoa, could hardly have exceeded $10^{-8}$ of the mass of the entire atmosphere ($5.14 \cdot 10^{21}$ g).

Among our other *qualitative* conclusions, which we believe worth emphasizing here, are the following:

No significant anomalies in the *global* radiation--i.e., the total downward flux of solar energy through a horizontal surface composed of the direct plus the diffuse radiation from the whole sky--clearly

attributable to the volcanic dust layers have been demonstrated, although definite diminutions in the direct (unscattered) sunlight are evident in the records. This does not contradict theoretical expectations from present knowledge of the effects of moderate turbidity on the pure molecular atmosphere, considering the forward scattering effects. Likewise, we have little reason to believe that the planetary albedo of the earth was significantly altered by the dust. It follows that the radiation balance at the boundaries of the atmosphere cannot have been significantly disturbed, so far as we may judge, by any of the three volcanic events here considered. However, it is quite likely that the broad spectrum (and polarization) of the daylight sky was variously altered in each case. This may have affected the photosynthetic process, for example, which is known to be sensitive to the skylight spectrum; but we are unable to estimate the magnitude of such effects, if any, in the absence of data.

We have found no evidence for a possible separation of the volcanic dust loading into tropospheric and stratospheric components nor any corresponding differences in composition, size distribution, and concentration, plausible as they may be. We feel that some of the existing ideas in this regard may be more inspired by the traditional meteorological notion of a tropopause as a true interface rather than by observation or theoretical expectation. Thus, the fact that Grams and Fiocco [1967] failed to detect any significant lowering in height of the stratospheric aerosol maximum with increasing latitude indicates that the responsible physico-chemical mechanism is not related to that of the tropopause. At any rate, our above estimates of the magnitude of the volcanic turbidity anomalies apply to the whole atmosphere rather than to any part thereof.

As to the specific particle size and size-distribution characteristic of each volcanic event, our present analysis shows that the Katmai and Agung particles must have been rather small with a relatively wide, continuous distribution, since no unusual blue-sun or Bishop's-ring phenomena were evident. We may assume that in both cases the particles were mainly generated *in situ* by the same processes that are responsible for the creation of "normal" stratospheric aerosols,

except for the higher number density resulting from the volcanic gaseous emissions. In the case of Krakatoa, on the other hand, we believe that the size distribution must have contained a second, narrow distribution around a large characteristic size, in addition to a long-lived, fine-particle component similar to that of the other volcanoes in size distribution. The larger component needed to explain the unusual optical phenomena noted may well have consisted of true particulates injected by the volcano directly into the stratosphere where they may have resided for as long as a year before settling. (The lifetime of the finer component may have exceeded five years in all three cases. However, this inference is rather tentative in view of possible interim injections from unknown volcanoes or other sources.)

Note also that although Krakatoa and Agung are both situated in the same latitude and longitude zone, the extent and mode of spreading of their respective dust material seem to have been quite different. This may be partly attributed to the different time of year in which the events occurred (August and March, respectively), and hence to possible differences in the pertinent large-scale motions; and perhaps also to the 80-year lapse between the two events, so that secular changes in the so-called general circulation may not be entirely discounted.

Our survey also shows that it is essential to institute reliable mechanisms for a continuous monitoring of turbidity anomalies produced by future volcanic activity, as well as of other long-period trends related to natural or man-made sources. We favor systems of proven value and operational facility, e.g., the re-establishment of permanent stations in strategic surface locations around the earth to record the narrow-band transmission of the atmosphere for sunlight and starlight [Deirmendjian, 1955a; Hodge, 1971]. (Obvious improvements using modern sensors and automation would much reduce the size, cost, and operating time of the old Smithsonian Observatory's system.) Such data could be supplemented by routine measurements of the aureole around the sun at fixed angular distances in order to better evaluate the physical nature of the particulates. On the other hand, we question the usefulness of the monitoring systems proposed recently

by an *ad hoc* study group [SCEP, 1970, pp. 200-202]. They favor the use of satellites to detect turbidity changes from albedo measurements (in our opinion the *least* reliable method, due to problems of interpretation) and a network of "Volz photometers" (a simplistic and unproven notion probably promoted by F. E. Volz, its presumed inventor; the specificity of the recommendation, to the point of naming a particular instrument without further elaboration, coming from a deliberative group of independent scientists, *ipso facto* casts doubts on its objectivity).

In addition, propulsion plants designed for stratospheric commercial transportation systems should be subjected to extensive ground testing prior to their operational use by objective, independent laboratories under simulated operational environments in order to determine the nature and amount of exhaust particulates.

A most important objective of a good monitoring and data reduction system is, of course, the unambiguous identification of the type and origin of the various components of the observed turbidity, such as volcanic, surface, man-made, as a function of geographical location, time of year, etc.

Finally, we come to the question of climatic changes related to turbidity anomalies and their simulation in the so-called numerical general circulation models of the global atmosphere. Clearly, the present survey indicates--quite apart from the difficulty of including such effects explicitly in existing programs--that it is premature to propose specific numerical "experiments" in this area. At least until more definite information becomes available, both on measurable climatic effects directly related to turbidity anomalies and on the role of various types and amounts of particulates in the radiation balance at the atmospheric boundaries (i.e., more complete theoretical results than presently available), it seems to us that any but the simplest experiment would be meaningless and hence useless. As a first stage in simplicity, we suggest a mere reduction, say by 10 or 20 percent, in the solar energy incident at the "top" of the model atmosphere, to simulate either a corresponding change in the solar constant or the action of some hypothetical absorbing layer outside the atmosphere. Obviously, such a model will not simulate real

volcanic dust--which we found to be a good scatterer within the atmosphere--but rather something like Fred Hoyle's [1957] *black cloud* interposed in space between the Earth and the Sun. However, the experiment does have the virtue of representing a *first-order effect* of turbidity and may help in our understanding of higher-order effects.

At such a time when the needed climatological data and theoretical knowledge, mentioned above, become available, we believe that the quantitative estimates of recent volcanically produced turbidity anomalies presented here, together with some appropriate time constant and the proposed polydisperse scattering models, should provide a rather good starting point for the introduction of turbidity effects into theoretical-quantitative models of climate modification.

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## APPENDIX

### Memorandum on the Astrophysical Observatory of the Smithsonian Institution*

by D. Deirmendjian

#### 1. Introduction

The remarks and suggestions that follow are motivated by the announcement of a reorganization of the Astrophysical Observatory of the Smithsonian under the direction of Prof. Whipple. It is hoped that they will assist the new director in increasing the scientific utility of the Observatory by enlarging its functions so as to cover a wider field of research with little additional cost and effort.

When Langley founded the Astrophysical Observatory in 1890, his main purpose appears to have been to detect daily and other periodic fluctuations in the solar constant and to try to correlate these directly to fluctuations in the weather. He invented the instrument (the bolometer) and the method of reducing the observations to extraterrestrial values (the Bouguer-Langley method) necessary to carry out the above program. From the latest volume of the Annals of the Astrophysical Observatory, (vol. 7, by L. B. Aldrich and W. H. Hoover, Washington, D. C. 1954) it would seem that Langley's original program and methods have been faithfully adhered to during the past fifty years with little modification.

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*Copy of an unpublished memorandum attached to a letter from Z. Sekera to Fred L. Whipple, dated 17 August 1955 (reproduced by permission from Z. Sekera).

The rapid advances both in instrumentation and theoretical investigations concerned with the interaction of sunlight with the atmosphere, achieved during the last few decades, as well as the progress in our knowledge of geophysical and astrophysical variables, justify a substantial revision of the original program of the Astrophysical Observatory, a revision which is long overdue.

## 2. Some Remarks on the Work Performed to Date

Thus far the criteria in making solar observations were set up approximately on the following premises: (a) The sun being the more or less variable main source of energy producing atmospheric disturbances, there must exist a direct link between these disturbances and the *total amount* of solar energy (called the solar constant) entering the confines of the atmosphere. (b) The atmosphere may be treated as a *fixed filter* or *absorbing layer*, so that its transmission characteristics may be estimated by solar photometry on the basis of very simple assumptions. Since the transmission is supposed to be given by each particular observation, there is little chance of erroneous extrapolation to the top of the atmosphere. Thus the sun's extraterrestrial energy can easily be found by ground observations. (c) The reliability of the measurements may be enhanced by using observatories on high mountains in desert country, and by taking observations on very "clear" days only, thus eliminating the interference by haze and humidity. To make sure that any observed periodicities of extraterrestrial origin are real it is desirable to establish observatories in both hemispheres and geographically as far distant from each other as possible. (d) Radiation can be accurately measured by a bolometer, based on the principle of black-body absorption and the thermal change of electrical resistance.

The above premises either are no longer valid or must be modified in line with more up-to-date ideas and available methods. In particular:

(a) Repeated attempts to correlate periodicities in the weather with supposed periodicities in the solar constant have failed to show any direct relation in a convincing manner. This kind of cause and effect relation is either too indirect to be detectable in terms of weather or is nonexistent. Furthermore it is possible that any solar weather control is more apt to be operative via some particular wave bands in the weakest part of the solar energy spectrum, than the total energy represented by the solar constant. It is not at all unreasonable to assume that very small effects brought about by gaseous absorption of short ultraviolet solar radiation in the uppermost layers of the atmosphere, may trigger off mechanisms which may result in larger disturbances in the lower atmosphere. Very little is known as yet about changes in the short ultraviolet energy output of the sun (see below).

As far as periodicities in the solar constant are concerned, the latest attempt at analyzing which appeared as recently as 1952 (cf. Smithso. Misc. Collections, vol. 117, No. 10, pp. 1-31), these are not significant enough. Indeed, they are mainly based on the assumption that the solar constant as obtained by the Smithsonian method is accurate to within one per mil. The 23 alleged periodicities with periods which are aliquot parts of the fundamental period found of 22.75 years, even if real, can hardly ever be correlated with weather features.

While it is difficult to visualize the significance or applications of the periodicities in the total energy from the sun, a detailed knowledge of the spectral energy distribution or extraterrestrial flux of solar energy at narrow wavelength intervals may be useful. The reason for this is that our knowledge regarding the composition of the atmosphere to high levels is increasing rapidly, especially after the war and with the perfection of rocket probing techniques. This means a better knowledge of the atmospheric absorption bands. The atmosphere as a scattering medium is also receiving more attention currently. The investigations are oriented in such a way that once the non-gaseous or turbidity components of the atmosphere and the nature of land and sea reflection

are known, and provided the sun's extraterrestrial energy *spectrum* is accurately known, it will be possible to know the radiation field in any direction within the atmosphere, as well as the spectral distribution of the energy flux from sun and sky (global radiation) received by unit horizontal area at the earth's surface (cf. *Nature*, 175, 459 (1955)). The importance of this physical quantity can hardly be over emphasized.

However, the observations of the Smithsonian Institution have not concentrated on this aspect but rather on the solar energy integrated over all wavelengths. The sun's energy spectrum is not regularly tabulated in the various volumes of the Annals but it can be obtained only as a by product of the so-called "long method." The latter consists in obtaining the monochromatic transmission coefficients for a rather limited number of fixed wavelengths, covering the spectrum between 350 and 1500 mμ. An ultraviolet and infrared correction is then added to obtain the absolute energy registered by the pyrheliometer, following more or less empirical criteria for their adjustment.

(b) The terrestrial atmosphere cannot be treated as a simple filter or absorbing medium for the incident solar radiation. It is predominantly a *scattering medium* with respect to the greater portion of solar radiation, with absorption taking place in more or less narrow bands and in more or less restricted regions of the atmosphere. Moreover the scattering is not given by a unique law (such as the Rayleigh law) but by several laws, depending on the size and character of the actual atmospheric particles. Nor is the atmosphere a fixed scattering and absorbing medium but it is rather quite variable. These time variations in the optical properties of the atmosphere are far more important than any periodicities in the solar constant, as far as the conversion of solar energy to other types manifested in the form of weather is concerned.

In particular, Langley's method does not always allow an accurate extrapolation to zero air mass, especially in the shorter

wavelengths (cf. *J. Opt. Soc. Amer.*, 43, 1158 (1953) and *J. Opt. Soc. Amer.** in preparation). The assumption that all energy scattered out by matter in the path of the parallel light is lost and not received at the bolometer slit is not correct. It can be shown that light scattered in the forward direction, even in a minute cone, is important enough in the presence of certain atmospheric scatterers, to render the Langley method inaccurate.

(c) An observatory location on a high mountain and limitation of observations to very "clear" days does not necessarily guarantee reliable solar constant values. It is almost certain that even above such mountain levels there exist more or less variable layers of what might be called natural aerosols (cf. *J. Opt. Soc. Amer.** in preparation). Failure to take these into account may introduce small errors. However it is true that Langley's method improves in accuracy the thinner the atmospheric layer traversed by sunlight. The reason why measurements should not be limited to very "clear" days and to high observatories is another one and will be made clear in section 3 below.

(d) Despite the development of very accurate and sensitive photometric instruments based on the photoelectric effect and cascade photo-multiplier, the Smithsonian observations continue to rely on the old spectrobolometer used without essential changes for the past forty years. There may be some virtue in the continuity of data obtained over long periods with a standard instrument, and this of course was the aim of the Astrophysical Observatory. However, this is hardly justification enough for the neglect to take advantage of more modern and efficient instrumentation and to modify the program of the observatory from time to time in line with new developments in scientific thought.

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*Subsequently published in its vol. 46, 565 (1956).

3. <u>Suggestions for Changes in the Future Work</u>

With the gradual perfection of rocket borne instruments and perhaps of future instrument carriers capable of taking measurements outside the earth's atmosphere it is to be expected that the exact nature of the solar energy before penetration into our atmosphere will be known in the near future. Any methods developed to get this information indirectly by measurements from within the atmosphere will then become obsolete. Nevertheless, the spectrophotometry of direct sunlight and diffuse skylight from the earth's surface still remains a valuable tool of research, not so much for solar physics but rather for atmospheric physics. In fact recent research indicates that the monochromatic optical thickness of the atmosphere as well as the skylight intensities very near the sun's disc (aureole) are very sensitive to changes in the size distribution of particles, especially of a diameter comparable to the wavelength in the visible spectrum. (These are the particles which act as the initial condensation nuclei in the formation of clouds.) A theory giving the scattering characteristics of a single spherical dielectric particle of this size was published some fifty years ago, while a theory solving the problem of radiative transfer in an atmosphere containing such particles in known quantities is currently being developed (cf. "Investigation of Polarization of Skylight" A.F.C.R.C. Contract AF 19(122)-239, Final Report Dept. of Meteorology, U.C.L.A. (1955)).

If the Astrophysical Observatory is to continue systematic solar observations, in view of the foregoing remarks and assuming that atmospheric physics may be considered within the scope of the Observatory's work, the following suggestions are put forward:

1. <u>Spectrophotometer and Instrumentation</u>: Replace the bolometer by photoelectric tubes and photomultipliers capable of measuring very weak radiation in as narrow

spectral bands as possible. The photometer slit should be as narrow as possible and its effective solid angle accurately known. Reflections of the sunlight before reaching the photosensitive surface should be avoided or kept at a minimum. In addition, an instrument (such as Dr. Evans' Harvard photometer) should be perfected, capable of measuring the monochromatic sky brightness of a known portion of sky very near the sun's limb. This or a similar photometer should be capable of measuring skylight coming from other directions also. If possible an instrument should be developed which can be used in measuring monochromatic fluxes of light coming from the whole sky, or well defined large portions of sky, in order to get the total sky radiation through a horizontal surface.

2. Location of Observatories: It is important to have observatories both near large oceans and in the midst of large continents for there are indications that the permanent turbid components of the atmosphere depend on the continentality or marine origin of the air. If the maintenance of a South American observatory is justified, there should be a station in the middle of the continent in addition to the one near the Pacific coast. Furthermore, observations which can be taken near sea level and to windward of an existing high mountain observatory may be very useful in ascertaining the optical properties of the intervening layer.

3. Choice of Days for Observation: Observations should not be limited to the clearest days but rather should be continuous whenever thick and complete cloud cover does not prevail. Observations should be particularly concerned with days when unusual atmospheric optical

phenomena occur, such as caused by invisible cirro-stratus, volcanic dust, smoke from forest fires, etc.

4. Presentation of Data: The observational data should be published so as to be of most use in the research on atmospheric physics by means of solar photometry. Interpolations, extrapolations, averaging over several days, statistically analyzed data, etc. should be avoided as far as possible. The original data as obtained at certain definite wavelengths and at the appropriate elevation of the sun should be clearly exhibited, together with the units of measurement, instrumental constants, state of the sky, barometric pressure and humidity, and any other pertinent data rendering the observations most usable to the researcher.